Abstract

Comparing compositional models of the terrestrial planets provides insights into physicochemical processes that produced planet-scale similarities and differences. The widely accepted compositional model for Mars assumes Mn and more refractory elements are in CI chondrite proportions in the planet, including Fe, Mg, and Si, which along with O make up >90% the mass of Mars. Here we present an alternative model composition for Mars that avoids such an assumption and is based on data from Martian meteorites and spacecraft observations. Our modeling method was previously applied to predict the Earth’s composition. The model establishes the absolute abundances of refractory lithophile elements in the bulk silicate Mars (BSM) at 2.26 times higher than that in CI carbonaceous chondrites. Relative to this chondritic composition, Mars has a systematic depletion in moderately volatile lithophile elements as a function of their condensation temperature. Given this finding, we constrain the abundances of siderophile and chalcophile elements in the bulk Mars and its core. The Martian volatility trend is consistent with \(<7\) wt\% S in its core, which is significantly lower than that assumed in most core models (i.e., \(>10\) wt\% S). Occurrence of ringwoodite at the Martian core-mantle boundary might have contributed partitioning of O and H into the Martian core.
1 Introduction

Mars is the second best-known planet in our solar system, given multiple space missions and cosmochemical studies on martian meteorites (McSween Jr and McLennan, 2014). Therefore comparison of physical and chemical properties of Mars with those of the Earth can provide important insights into the origin and evolution of the rocky planets, especially conditions for a habitable planet formation. Radioisotope dating of Martian meteorites demonstrates that its accretion and evolution occurred earlier than that of the Earth (Dauphas and Pourmand, 2011; Kruijer et al., 2017b; Bouvier et al., 2018). The rapid formation of Mars is consistent with a pebble accretion model for its formation, which predicts efficient accretion of Mars-sized planetary embryos before the dissipation of the protoplanetary disk (Johansen et al., 2015; Levison et al., 2015). Comparing the composition of Mars and the Earth (McDonough and Sun, 1995; McDonough, 2014) will provide insights into processes of planetary formation and evolution.

Compositional modeling of terrestrial planets requires determining the abundances and distribution of elements, given limited chemical data from their silicate shell, knowledge of the behavior of elements in different P-T-Xi-\(fO_2\) conditions (Table 1), and constraints from their geodetic properties. A compositional model for the bulk planet and its core and mantle can be used to understand the many and markedly different processes involved in its accretion and differentiation. Models for the chemical composition of Mars (e.g., Morgan and Anders, 1979; Longhi et al., 1992; Wänke and Dreibus, 1994; Lodders and Fegley Jr, 1997; Sanloup et al., 1999; Halliday et al., 2001; Burbine and O’Brien, 2004; Taylor, 2013) have been reviewed recently by Taylor (2013). Limited cosmochemical constraints and a lack of seismic data from Mars make it difficult to evaluate critically these competing models. Importantly, most existing models assume Mars’ major element composition equates to that in CI carbonaceous chondrites. Chondrites are undifferentiated assemblage of metal and silicates (Scott and Krot, 2014). Chondritic meteorites, especially the CI carbonaceous chondrites, are chemically similar to the solar photosphere (e.g., Palme et al., 2014), which is taken to reflect the Sun’s abundances of non-gaseous elements. At >99% of mass of the solar system, understanding the Sun’s composition and that of chondrites, the building blocks of the terrestrial planets, is key to understanding the sources and processes involved in making the planets.

The Wänke and Dreibus family of models (Wänke, 1981, 1987; Wänke and Dreibus, 1988, 1994; Dreibus and Wänke, 1984, 1987; Wänke and Dreibus, 1994) is the most widely accepted compositional model of Mars. It is based on chemical composition of Martian meteorites and assumes that Mn and more refractory elements (Table 1) are in CI-like proportions in Mars. Abundances of other less refractory elements are determined from chemical correlations with refractory or major elements. Many studies use a similar approach (Longhi et al., 1992; Halliday et al., 2001; Taylor, 2013; Taylor, 2013) revisited and updated the Wänke and Dreibus model using the more abundant, recent chemical data for Martian meteorites and spacecraft observations, and found no significant difference with the Wänke and Dreibus model. This model is the standard for most geochemical and geophysical modeling (e.g., Sohl and Spohn, 1997; Khan et al., 2018) and experimental works (e.g., Bertka and Fei, 1997; 1998a).
Here we present an alternative compositional model for Mars. We avoid the CI chondrite assumption. We base our model on data from Martian meteorites and spacecraft observations and use a method that was previously applied to predict the Earth’s composition. We determine a unique composition for the bulk silicate Mars (BSM) and a best fit, non-unique model for its core composition. By establishing the systematic depletion in volatile element in the BSM, we show that the core has \( \leq 7 \) wt\% S along with O and H as light elements. We discuss similarities and differences between the Earth and Mars and possible causes of these differences in a companion paper (Yoshizaki and McDonough, submitted).

2 Recent developments in understanding the solar system

Over the last decade technological advances and insights have revealed markedly new perspectives about the Sun’s composition and restrictions in the radial distribution of certain chondritic materials. Multiple challenges have been advanced regarding the solar photosphere’s composition, weakening the use of CI chondrites as a proxy for the bulk solar composition. Spectroscopic observations of the solar photosphere and assumptions about local thermodynamic equilibrium in modeling the photosphere’s composition are used to constrain its elemental abundances (e.g., Asplund et al., 2009). The solar metallicity problem notes the significant difference in estimates of the sun’s metallicity \( (Z_\odot, \text{abundance of elements in the Sun heavier than He}) \) from spectroscopic observations versus helioseismology (e.g., Basu and Antia, 2008; Haxton et al., 2013; Bergemann and Serenelli, 2014). The former method suggests 30 to 40% lower metal content in the Sun and the finding is at 5\( \sigma \) to 15\( \sigma \) outside the limits set by helioseismology for the Sun’s surface He abundance, the sound speed in the convective zone, and the depth of the convective zone boundary. Data from experiments on the opacity of metals in high temperature plasma (Bailey et al., 2015; Nagayama et al., 2019), the composition of solar wind particles (Schmelz et al., 2012), and measurements of solar neutrino flux (Haxton et al., 2013; Agostini et al., 2018) are in mutual agreement with findings from helioseismology, regarding the Sun’s metallicity. One solution to the problem is to have a significant increase in the Sun’s abundance of Mg, Si, S and Fe, which leads to a hotter core temperature (Basu and Antia, 2008; Asplund et al., 2009; Bergemann and Serenelli, 2014).

It is also important to recognize the accretion settings of different chondritic parent bodies. Isotopic distinctions (e.g., O, Ni, Cr, Ti, Mo, W) are now clearly established for the non-carbonaceous and carbonaceous (NC and CC, respectively) meteorite groups, including the chondrites (Warren, 2011; Dauphas and Schauble, 2016; Kruijer et al., 2017a). These differences likely originated because of a limited radial transport in the accretion disk, which may have been controlled by an early-formed young Jupiter (Walsh et al., 2011; Kruijer et al., 2017a; Raymond and Izidoro, 2017; Desch et al., 2018). As members of the inner solar system NC group, Mars and Earth are isotopically most similar to ordinary and enstatite chondrites, respectively, whereas CI chondrites are isotopically a part of the CC group meteorites, which are taken as sampling the outer solar system building blocks. Trace element chemistry of NC and CC meteorites supports this isotopic divide of two groups (Dauphas and Pourmand, 2015; Barrat et al., 2016). Thus, we recommend not using
a CI-chondrite compositional model for the inner terrestrial planets, including Mars. Here we develop a compositional model for Mars that is based on Martian rocks and is free of the CI chondrite assumption.

3 Data

A compilation of chemical and isotopic data of Martian meteorites was used in this study, with most data coming from the Martian Meteorite Compendium (Righter, 2017) and the online Met-Base database (https://metbase.org/). Shergottites, especially lherzolitic ones, were used to estimate composition of the bulk silicate Mars (BSM). Data for olivine shergottites (SHE-OI) and basaltic shergottites (SHE-B) provide robust compositional correlations. In most cases we excluded nakhlites, chassignites and other ungrouped Martian meteorites (e.g., Allan Hills (ALH) 84001) from our analyses because these samples are cumulates, metamorphosed cumulates, or more complex lithologies rather than simple melt-derived rocks (McSween Jr, 2008). Paired Martian meteorites (e.g., Northwest Africa (NWA) 2975 and NWA 2986) are treated as one sample. We exclude NWA 7397 lherzolitic shergottite from our analyses because it has experienced a complicated formation processes (Howarth et al., 2014). Several analyses which are unrepresentative due to sample heterogeneity or terrestrial contamination are also excluded from our dataset. Average values of elemental abundances in each Martian meteorite are calculated and used in the compositional modeling. We also use data from the Martian surface as measured by spacecraft missions using a gamma-ray spectrometer (GRS) (Boynton et al., 2008; Taylor et al., 2006a,b). For some elemental abundances, we adopted estimates by recent studies (e.g., Yang et al., 2015; Wang and Becker, 2017; Tait and Day, 2018). Errors are reported as 1 standard deviations, unless otherwise noted. For a CI chondritic composition, we adopted the value proposed by Palme et al. (2014) for most elements, with some modifications for halogens (Clay et al., 2017), Mo, Tl, Bi (Wang et al., 2015), highly siderophile elements (Day et al., 2016), and U (Wipperfurth et al., 2018) (Table 2).

4 Composition of the bulk silicate Mars

4.1 Refractory lithophile elements

Refractory lithophile elements (RLE; Table[1]) remain in the silicate shell during core-mantle differentiation and their relative abundances show limited variation (generally <10%) among chondritic meteorites (e.g., Masuda, 1957; Coryell et al., 1963; Larimer and Wasson, 1988; Wasson and Kallemeyn, 1988; Bouvier et al., 2008b). Thus, if you establish the absolute concentration of one RLE, you can calculate the abundances of all of the others based on chondritic ratios (e.g., Sm/Nd, Lu/Hf, Ca/Al).

As a first step in determining the composition of Mars, we independently tested if Mars has chondritic ratios of RLE. We used ratio-ratio plots to demonstrate that the trends cross at the in-
intersection of chondritic values when using 4 independent refractory lithophile elements (Figure 1). These chemical trends see through previous melt-residue differentiation events back to their primordial compositions, which are chondritic RLE ratios. Thus, these chemical trends document negligible (within uncertainties) fractionation of RLE in the undifferentiated BSM composition. Given this finding, all RLE are taken to be in chondritic relative proportions in the BSM and, as a follow on, in the bulk Mars too. From this starting point, we can directly determine the BSM abundances of most elements, whose concentrations are correlated with ratios or abundances of refractory lithophile elements in Martian rocks, due to chemical trends resulting from melt-residue differentiation.

The absolute abundance of the RLE are established using variation diagrams involving a single RLE versus an RLE ratio (Figure 2). The lherzolitic shergottite (SHE-L) are the best recorders of melt depletion trends, providing an accurate estimate of the primitive BSM composition. Using these melt depletion trends for multiple element combinations, we estimate the absolute abundance of the refractory lithophile elements in the BSM at $\sim 2.26$ times that in CI chondrites with $\sim 10\%$ uncertainties (cf. $2.75 \times$ CI in the BSE; McDonough and Sun, 1995).

4.2 Major elements (Mg, Si, Fe)

Magnesia correlates negatively with RLE abundances in shergottites (Figure 3) and trends for multiple RLE establish the MgO abundance at 31.0 $\pm$ 2.0 wt% for the BSM. There is limited variation in silica contents in shergottites (Figure 4), reflecting silicon’s bulk distribution coefficient of $\sim 1$ during silicate melt production. The SiO$_2$ content of the Martian surface as measured by gamma-ray spectroscopy (GRS) (Boynton et al., 2007) overlaps with the range seen in basaltic shergottites, confirming the bulk crust of Mars is basaltic (e.g., Taylor and McLennan, 2009; McSween Jr and McLennan, 2014 and references therein). By averaging the SiO$_2$ abundances in shergottites, we estimate 45.5 $\pm$ 1.8 wt% SiO$_2$ for the BSM. Assuming no Si or Mg in the Martian core, our BSM and bulk Mars model compositions have a Mg/Si value of 0.88 $\pm$ 0.07, which agrees with an estimate based on Si isotope systematics (0.86 $\pm$ 0.05; Dauphas et al., 2015). Uncertainty in Mg/Si value for the BSM does, however, overlap with average values for ordinary (~0.82) and carbonaceous chondrites (~0.92).

As compared to the Earth, Mars is more oxidized and Martian meteorites, including the least evolved samples, have distinctly lower Mg# (atomic ratio of Mg/(Mg + Fe)) relative to basalts from the Earth (e.g., Wadhwa, 2001, 2008; Herd et al., 2002). Estimates of the Mg# of the Martian mantle range between 0.7 and 0.8 (Table 3). Agee and Draper (2004) argued that a bulk Martian mantle or shergottite source region has a Mg# closer to that of H-type ordinary chondrites (0.79–0.82; Rubin et al., 1988). Petrological and geochemical data support the H chondrite like Mg# of the Martian mantle (Minitti et al., 2006; McCoy et al., 2016; Draper et al., 2005; White et al., 2006; Borg and Draper, 2003). Adopting an Mg# of 0.79 for the Martian mantle gives an FeO content of
14.7 ± 1.0 wt% in the BSM. Collectively, these findings for the RLE (Al, Ca and Ti), magnesia, silica and ferrous iron sets the BSM composition at a total of 97.8 wt% (Table 4).

4.3 Non-refractory lithophile elements

4.3.1 Manganese, chromium and vanadium

The MnO concentration in the BSM (Wänke and Dreibus, 1994; Taylor, 2013) is established foremost from the nearly constant FeO/MnO ratio in Martian meteorites (shergottite average 39.4 ± 0.07) and the Martian mantle’s FeO content, producing a BSM having 0.37 ± 0.07 wt% MnO. An alternative approach uses experimental studies on the partitioning of Mn at 1–2 GPa during melting (Baratoux et al., 2011; Filiberto and Dasgupta, 2011). The measured olivine/melt partition coefficient for MnO (Takahashi and Kushiro, 1983; Walter, 1998; Herzberg and Zhang, 1996; Wasylenki et al., 2003; Le Roux et al., 2011) is 0.93 ± 0.04 (Taylor, 2013). Using this value and a shergottite average MnO content (0.48 ± 0.06 wt%) predicts 0.44 ± 0.06 wt% MnO in the BSM. The MnO content of the BSM obtained using these methods are in agreement within errors.

Chromium correlates with Al ($R^2 = 0.72$) in lherzolitic and olivine-phyric shergottites. Where the Cr-Al trend crosses the Al content of the BSM yields 0.88 ± 0.15 wt% Cr$_2$O$_3$ in the BSM. The CI-normalized abundances in the BSM for Cr and major elements (Mg, Si and Fe), with similar condensation temperatures (Lodders, 2003), is used to conclude that the Martian inventory of Cr is hosted solely in the mantle.

In the solar nebula vanadium behaves as a refractory lithophile element (Lodders, 2003). During the formation of the Earth’s core it was equally lithophile and siderophile in its behavior under high P-T and/or reducing condition (Ringwood and Hibberson, 1990; Wade and Wood, 2005; Wood et al., 2006, 2008; Corgne et al., 2008; Siebert et al., 2013), resulting in half of the Earth’s inventory of V to be in the core (McDonough, 2014). Estimates of the V content of the BSM using correlation diagrams yields ∼130 ppm V, which is an equivalent concentration to that of the other RLE (∼2.26 × CI) (note, here and throughout the paper ppm and ppb will refer to parts per million and billion by weight, respectively). Thus, we conclude that V behaved exclusively as a lithophile element during Mars’ core formation and is wholly concentrated in the BSM.

4.3.2 Sodium and potassium

A log-log abundance plot of Na and Al in lherzolitic and olivine-phyric shergottites show a well-defined slope ∼1 correlation ($R^2 = 0.83$), indicating nearly equal incompatibility of these elements during a partial melting of the Martian mantle. The limited variation in Na/Al ratio in shergottites (0.44 ± 0.10) is consistent with a Na$_2$O content of 0.59 ± 0.13 wt% in the BSM.
The radioactive elements, K, Th and U, have similar partitioning behavior during mantle melting. Consequently, the Martian surface K/Th value of 5300 ± 220, measured by Gamma Ray Spectroscopy (GRS) aboard Mars Odyssey (Taylor et al., 2006a,b), is taken as the bulk K/Th value. Both Th and U are RLE and this K/Th value gives 0.043 ± 0.005 wt% K₂O in the BSM. Our BSM model has Na/K of 12 ± 3, comparable to the Earth’s ratio (11) and overlapping within errors of the lower chondritic value (9.1 ± 1.3; Wasson and Kallemeyn, 1988). Importantly, this ratio of elements increases as a function of the relative condensation temperature, such that Na/K increases from CI to CM to CO/CV chondrites as refractory to volatile element ratios increase (Wasson and Kallemeyn, 1988). This trend is consistent with higher Na/K in the Earth and Mars.

4.3.3 Rubidium

Martian meteorites show a negative trend in a plot of initial ε₁⁴₃Nd vs ⁸⁷Sr/⁸⁶Sr values, establishing the Martian mantle array (Figure 5). The Martian mantle array is shifted to higher ⁸⁷Sr/⁸⁶Sr values compared to the Earth’s mantle array, consistent with Mars’ higher content of volatile elements. The Martian mantle array yields an initial ⁸⁷Sr/⁸⁶Sr value of 0.709–0.721 in the BSM assuming an initial ε₁⁴₃Nd = 0 for the planet. If Rb/Sr fractionation took place at the earliest stages of accretion, then Mars’ initial ⁸⁷Sr/⁸⁶Sr value is consistent with a Rb/Sr of 0.08 ± 0.04, and thus 1.5 ± 0.8 ppm Rb in the BSM. Analogous to Na/K, K/Rb ratio for Mars is 300, which is higher than the CI value (250) and not as volatile depleted as the Earth’s value (400).

Alternatively, a log-log concentration plot of Rb vs La (slope = 1.0 ± 0.1; R² = 0.84) reveals their similar partition coefficients during mantle melting. Using La (an RLE) abundance in the BSM, we predict 0.9 ± 0.4 ppm Rb in the BSM, comparable to that based on the Nd- and Sr-isotopic systematics. An average of these estimated values yields 1.2 ± 0.4 ppm Rb in the BSM.

4.3.4 Cesium

In a log-log concentration plot, Cs and La show a trend with a slope of 1.0 ± 0.1 (R² = 0.78). The Cs/La ratio in shergottites (0.14 ± 0.10) leads to 0.08 ± 0.06 ppm Cs in the BSM. This value is consistent with an average Rb/Cs ratio of shergottites (16.0 ± 3.0) which indicates 0.07 ± 0.03 ppm Cs in the BSM.

4.3.5 Lithium

Lithium correlates with Nb and LREE in Martian meteorites. Intercepts of the Li-RLE trends and the BSM abundances of the RLE (~2.26 × CI) yields Li concentration in the BSM of 1.8 ± 0.4 ppm.
4.3.6 Boron

Boron and Ca are well-correlated in lherzolitic and olivine-phyric shergottites, except for five high-B (>10 ppm B) olivine-phyric shergottite samples discussed by Day et al. (2018). We agree that these five high-B samples are possibly affected by terrestrial weathering (Curtis et al., 1980; Yang et al., 2015; Day et al., 2018) and exclude them from consideration. The average B/Ca (0.41 ± 0.25) in lherzolitic and olivine-phyric shergottites leads to 0.84 ± 0.53 ppm B in the BSM.

4.3.7 Gallium

Gallium and Al are positively correlated in shergottites \(R^2 = 0.84\). An average Ga/Al ratio for shergottites \((4.6 ± 0.9)\) yields 8.7 ± 1.9 ppm Ga in the BSM. Similarly, the Earth’s average Ga/Al ratio for basalts and mantle rocks is 4.3 (McDonough, 1990), the same as that for Mars. The abundance of Ga in the BSM and BSE (bulk silicate Earth) points to its lithophile behavior during core–mantle differentiation and these cores having negligible quantities of Ga.

4.3.8 Halogens

Efforts to estimate the abundances of the halogens (Cl, F, Br, I) are fraught with challenges, because halogens are fluid-mobile and thus readily lost during magma degassing and by alteration processes (e.g., Filiberto et al., 2019 and references therein). Many have attempted to filter Martian meteorite data affected by such secondary processes (Dreibus and Wänke, 1985; Dreibus and Wänke, 1987; Treiman, 2003; Filiberto and Treiman, 2009; Filiberto et al., 2016, 2019).

Filiberto and Treiman (2009) and Filiberto et al. (2016) constrained Cl/La value for the Martian basalt to be 51 ± 17, which is higher than the terrestrial Cl/La ratio (21 ± 6). This Cl/La ratio and the BSM abundance of La \((0.55 ± 0.05\ ppm)\) yields 28 ± 10 ppm Cl in the BSM, which is our preferred value. We observed a well-defined correlation between Cl and Sr in Martian meteorites \((R^2 = 0.79)\), which corresponds to 32 ± 21 ppm of Cl in the BSM. Although errors in the Cl abundance estimated by two methods are large, these values are consistent in general. Taylor et al. (2010) used GRS data for Cl/K observed a ratio of 1.3 ± 0.2, comparable to the Cl value (1.28), and yielded 0.5 wt% Cl in the BSM. However, others (Keller et al., 2006; Filiberto et al., 2019) conclude that Cl abundance in the Martian surface reflects secondary deposition and thus high Cl/K ratios of the Martian surface might not be an indicator of the primitive Cl content of the BSM.

There is limited F data for Martian meteorites. The few shergottites samples with coupled F and B data show a correlation \((R^2 = 0.80)\), indicating 16 ± 12 ppm F in the BSM. Filiberto et al. (2016) reported shergottites and terrestrial basalts with a mean Cl/F of 0.8 ± 0.4. Given terrestrial and Martian basalts are modified by magma degassing, which decreases the initial value, whereas alteration increases the ratio, Filiberto et al. (2016) suggested Cl/F ∼ 1 in the Martian mantle. Using our estimate of the BSM abundance for Cl, we obtained the BSM abundance for F of ∼28 ppm, our preferred value.
An average Cl/Br ratio in Martian meteorites (224 ± 140; Filiberto et al., 2016) indicates 0.12 ± 0.09 ppm Br in the BSM. The limited variation in Cl/Br values in Martian meteorites, chondrites and the BSE and similarly, albeit with even less data, for Cl/I in chondrites and the BSE, lead Clay et al. (2017) to conclude that the 50% condensation temperature of Cl is much lower than 948 K (Figure 6). Their conclusion has also been supported by newer thermodynamic calculations suggesting the equilibrium 50% condensation temperature for Cl is 472 K (Wood et al., 2019).

There is limited data for I in Martian meteorites, the most volatile of the lithophile elements (Table 1). We do not observe any clear correlation of the halides with other elements. Thus the I abundance in the BSM is defined by the Martian volatility trend (Figure 6).

4.4 Siderophile and chalcophile elements in the BSM

4.4.1 Refractory and major siderophile elements in the BSM

These elements include W and Mo along with Fe, Ni, and Co (Table 1). Nickel in lherzolitic and olivine-phyric shergottites correlate with Mg and their average Ni/Mg ratio is 19.4 ± 5.0, consistent with 0.046 ± 0.012 wt% NiO in the BSM. Likewise these rocks have an average Co/Ni ratio of 0.27 ± 0.10, consistent with 96 ± 44 ppm Co in the BSM.

Mo abundance in shergottites is highly variable between 0.05–0.7, and is not well correlated with lithophile elements. The origin of the wide variation in Mo contents in shergottites is not constrained. Yang et al. (2015) attributes this variation to hydrothermal processes, whereas Noll Jr et al. (1996) argues that Mo is not fluid-mobile. Based on a broad Mo-Ce co-variation, Righter and Chabot (2011) and Yang et al. (2015) estimated the BSM abundance of Mo is 0.08–0.6 ppm. We estimate the BSM to have between 0.1–0.8 ppm Mo.

In a log-log plot for all shergottites, W and Th shows a linear trend with a slope 1.1 ± 0.2, indicating their similar incompatibilities. Using an average W/Th ratio in shergottites (1.0 ± 0.5), W concentration in the BSM is estimated to be 0.07 ± 0.04 ppm.

4.4.2 Highly siderophile elements

Highly siderophile elements (HSE) include Re, Os, Ir, Pt, Ru, Rh, Pd and Au. Brandon et al. (2012) and Tait and Day (2018) estimated the BSM abundances of HSE based on co-variation between HSEs and MgO in shergottites. Although Re, Pd and Pt are poorly correlated with Mg, Tait and Day (2018) assumed MgO ~35 wt% in the BSM and obtained flat, chondrite-like HSE patterns in the BSM, with absolute abundances at (0.010 ± 0.003) × CI. The MgO composition of the BSM used by Tait and Day (2018) is slightly higher than our estimate (MgO ~ 31.0 wt%). However, this difference is negligible for the predicted abundances of the HSE in the BSM and so we adopt the Tait and Day (2018) estimate.
Similarly, Au also show a broad correlation with MgO composition in shergottite. The correlation indicates $\sim 2$ ppb Au in the BSM, which corresponds to the BSM abundance of $\sim 0.01 \times CI$ (Figure 6). Thus it is likely that the most BSM budget of Au also originated in the late accretion of materials with chondritic HSE composition.

There are limited data for Rh in Martian meteorites with a wide variation, and Rh is not correlated with other elements in Martian meteorites. If we take the mean Rh concentration in the Martian meteorites ($\sim 2$ ppb) as the BSM abundance of Rh, this value corresponds to $\sim 0.02 \times CI$ in the BSM, which is close to that of other HSE (Figure 6; Tait and Day, 2018).

### 4.4.3 Moderately volatile siderophile elements

Moderately-volatile, siderophile elements include P, Cu, Ge, As, Ag, Sb, Sn, Pb and Bi. A well-defined positive correlation for P vs Y ($R^2 = 0.88$), with a slope of $1.1 \pm 0.1$, is seen for shergottites. Their average P/Y value in shergottites ($0.021 \pm 0.005$) is used to estimate the $P_2O_5$ at $0.17 \pm 0.05$ wt% in the BSM.

Wang and Becker (2017) estimated Cu in the BSM abundance at $2.0 \pm 0.4$ ppm, based on Cu and MgO correlations in shergottites. Similarly, we find positive Cu vs Ti correlations in shergottites, except for some anomalously high Cu/Ti samples (Cu/Ti $< 0.005$), and their average Cu/Ti value ($0.0026 \pm 0.0005$) yields $2.6 \pm 0.6$ ppm Cu is in the BSM, which is comparable to the estimates by Wang and Becker (2017).

Germanium in Martian meteorites do not correlate with lithophile elements, as mantle melting produces little variation in Ge content in the melt and residue (Ringwood, 1966). Germanium contents in shergottites ranges from $0.8 \pm 0.4$ ppm in lherzolitic shergottites to $1.5 \pm 0.3$ ppm in olivine-phyric shergottites and $0.7–2.0$ ppm basaltic shergottites. From a weak negative correlation between Ge and MgO in lherzolitic and olivine-phyric shergottites, we estimate Ge abundance in the BSM is $0.6 \pm 0.4$ ppm.

In terrestrial rocks, As correlates with Ce (Noll Jr et al., 1996). In contrast, As in shergottites does not correlate with tested lithophile elements (Yang et al., 2015). Arsenic is a fluid-mobile element (Yang et al., 2015); thus, the lack of any As-Ce correlation in shergottites might be due to hydrothermal processes. Correcting for hydrothermal processes, Yang et al. (2015) suggested $30 \pm 25$ ppb As in the BSM.

We do not observe any correlation of Sb and Ag with lithophile elements. Yang et al. (2015) estimated the BSM abundance of Sb to be $0.01–0.03$ ppm using a roughly constrained Sb/Pr ratio of Martian meteorites and the BSM abundance of Pr of $0.17$ ppm from Lodders and Fegley Jr (1997). We accept their estimate given that $0.21$ ppm Pr in our model does not make significant changes in the estimation of the BSM abundance of Sb. A log-log plot of Sb versus Ag concentrations in
shergottites (slope \(\sim 1\)) indicates their similar compatibility during differentiation. The average
Sb/Ag ratio of 1.4 ± 1.0, with Ag being 0.02 ppm in the BSM with \(\sim 90\%\) uncertainty.

Yang et al. (2015) estimated the BSM abundance of Sn, Cd and In based on their broad correlation
with lithophile elements in Martian meteorites and a model of BSM abundances for lithophile
elements by Lodders and Fegley Jr (1997). The observed correlation resulted in more than a factor
of two uncertainties in their estimates. Although abundances of lithophile elements in our model is
different from that of Lodders and Fegley Jr (1997), this difference does not produce any significant
decrease in the estimated BSM abundances of Cd, In and Sn in Yang et al. (2015). Thus, we accept
the BSM abundances of these elements given by Yang et al. (2015).

The BSM abundance of Pb is constrained from U-Pb isotopic systematics of Martian mete-
orites. Martian meteorites have a wide ranges of \(\mu = \frac{^{238}U}{^{204}Pb}\) values and Pb isotopic hetero-
geneity (Nakamura et al., 1982; Misawa et al., 1997; Borg et al., 2005; Bouvier et al., 2005; 2008a,
2009; Chen and Wasserburg, 1986; Gaffney et al., 2007; Moriwaki et al., 2017; Bellucci et al., 2018).
The recent review of Martian \(\mu\) value (\(\sim 3.6\)), the average \(^{208}Pb/^{204}Pb\), \(^{207}Pb/^{204}Pb\) and
\(^{206}Pb/^{204}Pb\) in Martian meteorites of 33, 12.6 and 13, respectively, and 18 ppb U (Table 5) yields
0.26 ± 0.05 ppm Pb in the BSM.

Bismuth is correlated with Th in shergottites (except for high-Bi samples EETA 79001, Tissint
and NWA 5990) (Yang et al., 2015), which yields 2 ± 1 ppb Bi in the BSM.

4.4.4 Moderately volatile chalcophile elements

Moderately-volatile, chalcophile elements include Zn, Te, Se, S and Cd. As mentioned in Section 4.4.3, we adopt the BSM abundances of Cd estimated by Yang et al. (2015).

Zinc and Lu positively correlate in shergottites \((R^2 = 0.63)\) as does Zn and Ti \((R^2 = 0.56)\),
which yield 45 ± 15 ppm and 40 ± 15 ppm Zn, respectively, in the BSM. These two estimates
agree within errors.

Taylor (2013) estimated Zn in the BSM at 18.9 ± 1.5 ppm, whereas as Yang et al. (2015)
proposed 50–70 ppm. Taylor (2013) estimate was based on a Zn–Sc correlation, which is poorly
correlated in our dataset \((R^2 = 0.27)\). The relatively flat Zn–Sc trend observed by Taylor (2013)
leads to a poor control on an accuracy estimate. Yang et al. (2015) assumed an olivine-rich Martian
mantle and bulk mineral-melt partition coefficients for Zn of \(\sim 1\) during mantle melting (Le Roux
et al., 2011; Davis et al., 2013). The positive Zn-Lu correlation in shergottites reflects the slightly
incompatible nature of Zn in the Martian mantle. Thus, Yang et al. (2015) overestimates the Zn
abundance in the BSM.

There is a wide range of estimates for the abundances of S-Se-Te in the BSM (Franz et al.,
2019; Wang and Becker 2017 and references therein). The limited variation in S/Se/Te ratios,
indicating little degassing loss of S, Se and Te, which lead Wang and Becker [2017] to derive the BSM abundance of these elements at 360 ± 120 ppm, 100 ± 27 ppb and 0.50 ± 0.25 ppb, respectively. Here we adopt their estimates.

### 4.4.5 Volatile chalcophile elements

Volatile siderophile and chalcophile elements include In, Tl and Hg. As described in Section 4.4.3, we adopt the BSM abundances of In estimated by Yang et al. [2015] in our BSM model. Tl is positively correlated with Sm ($R^2 = 0.73$) in shergottites with a slope of 1.29 ± 0.38 in a log-log slope, indicating 4 ± 2 ppb Tl in the BSM.

Limited data exist for Hg ($\leq 0.7$ ppm) in Martian meteorites (Ehmann and Lovering [1967], Weinke [1978], Treiman and Lindstrom [1997]). It is readily contaminated by terrestrial sources, it is a highly volatile element, and it is lost during magma transport (e.g., Treiman and Lindstrom [1997]). Therefore it is very difficult to constrain the BSM abundance of Hg. Estimates of the solar abundance of Hg (few hundred ppb) are poorly constrained and Hg is undetectable in the solar photosphere (Lauretta et al. [1999], Grevesse et al. [2015], Meier et al. [2016]). The BSM abundance of Hg is estimated to be $\sim 7$ ppb, with uncertainty ranging from $\leq 1$ to a few tens of ppb.

### 4.5 Atmophile elements

Mars’ volatility trend (Figure 6) cannot provide strong constraints on the planetary abundances of atmophile elements (H, C, N, O, noble gases; Table 1), given their distinctive behaviors compared with less volatile elements that are retained in rocks. However, approximate estimates of these elements in the BSM were predicted from the Martian volatility trend. An extrapolation of the Martian volatility trend to the lower temperature indicates their abundances to be 0.001–0.01 × CI.

Based on the water contents of Martian meteorites and its constituting hydrous phases (e.g., apatite, amphibole, glass), the BSM is estimated to have $\sim 140$ ppm H$_2$O (e.g., McCubbin et al. [2016a,b], Filiberto et al. [2019]). This H$_2$O content corresponds to $\sim 16$ ppm H in the BSM, which is lower than that is expected based on the volatility trend. The low H abundance in the BSM might reflect H incorporation into the core (see Section 5.1) and/or loss of H from the mantle by degassing. Estimates for the water content of the Martian mantle range between 14–250 ppm, which likely reflects a heterogeneous water distribution in the mantle (e.g., McCubbin et al. [2010, 2016a,b]). The BSM might have contained $\sim 1000$ ppm H$_2$O if volatile-rich chondritic materials were added during a late accretion (i.e., 0.6–0.7% by mass of Mars), as indicated by highly siderophile element abundances and Os isotope systematics in shergottites (Tait and Day [2018]).

Filiberto et al. [2019] estimated the BSM abundances of C and N, based on mean C/H and N/H ratios in CI and CM carbonaceous chondrites (2.1 and 0.11, respectively; Alexander et al. [2019]).
and a BSM estimate for its H$_2$O content. This method assumes the BSM inventory of these atmophile elements is dominantly by a late addition component (e.g., Tait and Day 2018) with CI- and CM-chondritic chemical composition [Filiberto et al. 2019]. These assumptions lead to a BSM with ~140 ppm H$_2$O, ~32 ppm C and ~1.6 ppm N. These estimates have large uncertainties, given the degree of mantle degassing is essentially unconstrained.

5 Composition of the Martian core

5.1 Compositional model of the Martian core

To constrain the Martian core composition, we modeled geophysical properties of Mars. Mineralogy, radial density distribution and seismic velocity profiles in the Martian mantle were computed using the Gibbs free energy minimization method that is employed in the thermodynamic modeling code Perple_X version 6.8.6 (Connolly, 2009). Calculations for the BSM composition (i.e., primitive Martian mantle composition; Table 4) were performed using thermodynamic parameters of Stixrude and Lithgow-Bertelloni (2011) within a chemical system Na$_2$O–MgO–Al$_2$O$_3$–SiO$_2$–CaO–FeO. Temperature profile in the Martian mantle (areotherm) is estimated assuming an Earth-like profile [Katsura et al. 2010], mantle potential temperature of ~1,500 K (Baratoux et al. 2011, 2013; Putirka, 2016; Filiberto, 2017), a lithosphere thickness of 200 km (Grott et al. 2013) and conductive and adiabatic thermal gradients of 2.7 K/km and 0.12 K/km, respectively (Verhoeven et al. 2005), which are consistent with the surface heat flux estimates [Parro et al. 2017; Samuel et al. 2019]. Figure 7 shows a result of the thermodynamic modeling.

We also estimated mineralogy and physical properties of the present-day Martian mantle (not the BSM), which is calculated using compositional models for the BSM (Table 4) and the Martian crust (Taylor and McLennan, 2009), and a crustal mass fraction of ~5 % in the BSM. We observed small changes in the modal abundances of mineral species, but the density profiles are similar in the present-day and primitive mantle models. Thus, the use of the density profile obtained using the composition of primitive mantle or present-day mantle makes negligible changes in our discussion.

Using the obtained radial density profile in the Martian mantle, we computed the Martian interior structure. Here we consider Mars as a spherically symmetric body divided into three layers (crust, mantle and core). The average crustal thickness of 50 km, which is consistent with geophysical and geochemical constraints (Zuber et al. 2000; McGovern et al. 2002, 2004; Wieczorek and Zuber 2004; Ruiz et al. 2009), is adopted in the modeling. We consider a crust of basaltic composition (e.g., Taylor and McLennan, 2009; McSween Jr and McLennan, 2014 and references therein) with a density of 3010 kg/m$^3$, which is in agreement with gravity and topography studies [McKenzie et al. 2002; McGovern et al. 2002, 2004; Phillips et al. 2008].
of inertia (MOI) in Mars are expressed as a function of radial density distribution as:

\[ M = 4\pi \int_0^a \rho(r)r^2 dr \]  
\[ C = \frac{8\pi}{3} \int_0^a \rho(r)r^4 dr, \]  

where \( M \) is mass, \( C \) is MOI, \( a \) is Mars’ equatorial radius (3389.5 km; Seidelmann et al., 2002), \( \rho(r) \) is density and \( r \) is distance from the center of the planet.

Sulfur abundance in the Martian core has been of particular interests. Previous estimates vary between 3.5–25 wt% S (Table 6; also see Franz et al., 2019). With the Martian mantle having 360 ppm S (Wang and Becker, 2017), the Martian volatility curve restricts the core to having \( \leq 7 \) wt% S (Figure 8) if its core mass fraction is 18 wt% (see below). The low S composition for the Martian core is supported by metal-silicate partition coefficients for S (Rose-Weston et al., 2009; Boujibar et al., 2014; Wang and Becker, 2017) and an experimental study of Fe isotopic fractionation under high pressure and temperature (Shahar et al., 2015). If we adopt a higher S content for the BSM (up to 2000 ppm; Gaillard et al., 2013; Ding et al., 2015), the S content of the Martian core decreases. Thus, we conclude that the Martian core has \( \leq 7 \) wt% S.

This estimate stands in contrast to widely adopted Martian core models that argue for \( >10 \) wt% S (Wänke and Dreibus, 1994, Taylor, 2013). Such a high sulfur content for the bulk Mars would, in principle, treat S differently with respect to other moderately volatile elements. We find no justification for such a S enrichment (Figure 8).

For the Earth’s core, C, H, S, Si and O are proposed as candidate light elements that decrease its density (Birch, 1952, 1964). Given the P-T-fO\(_2\) condition of the Martian interior, we do not expect Si in the Martian core (Wade and Wood, 2005; Corgne et al., 2008), which is supported by a lack of Si isotopic fractionation in Martian meteorites resulted from a metal-silicate segregation (Zambardi et al., 2013; Dauphas et al., 2015). We also exclude C as a candidate light element in the Martian core since the Martian meteorites lack C isotope fractionation unlike terrestrial rocks (Grady et al., 2004; Wood et al., 2013) and addition of C does not efficiently decrease the core density (Wood, 1993; Bertka and Fei, 1998b). In contrast, experimental studies indicate that O and H can be incorporated into the Martian core (Okuchi, 1997; Shibazaki et al., 2009; Tsuno et al., 2011) and their addition can decrease the core density (Badding et al., 1992; Bertka and Fei, 1998b; Zharkov, 1996; Zharkov and Gudkova, 2005). Moreover, having the Martian core-mantle boundary in contact with ringwoodite can contribute to incorporation of H and O into the Martian core (Shibazaki et al., 2009; Tsuno et al., 2011; O’Rourke and Shim, 2018). Thus, we consider O and H in addition to S as candidates for light elements in the Martian core.

Densities of solid \( \gamma \)-Fe (Ahrens et al., 2002), FeS (Fei et al., 1995), Fe\(_3\)O\(_4\) (Reichmann and Jacobsen, 2004) and FeH (Badding et al., 1992) at core pressures and temperatures were calcu-
lated using a third-order finite strain Birch-Murnaghan equations of state (Stixrude and Lithgow-Bertelloni, 2005), following previous studies (e.g., Rivoldini et al., 2011; Khan et al., 2018). Temperature profile in the core is calculated based on the temperature of core-mantle boundary (Figure 7) and a convecting core.

By fitting mass, density and MOI in the three layers and the bulk planet to the geodetically constrained values (Table 7), the density and composition of the metallic core is estimated. Finally we obtained a core composition with 6.6 wt% S, 5 wt% O and 0.8 wt% H, and a mass fraction of 18% as our best estimate. This core model yields bulk planetary Fe/Si ratio of 1.36, which is lower than the CI value (1.74) but within the range of chondritic meteorite compositions (1.0–1.8; Wasson and Kallemeyn, 1988). With a mass fraction of 18%, the Martian core radius is 1580 km (i.e., 1810 km deep) and its mean density is ∼7060 kg/m³ (Figure 9).

Siderophile element abundances in the bulk Mars and its core were constrained by the BSM and core model (Table 8). This yields the Fe content in the core and bulk Mars to be 79.5 wt% and 23.7 wt%, respectively. Given chondritic meteorites (aside from the few examples of iron rich chondrites (e.g., CB)) show limited variation in Fe/Ni (∼17.4) and Ni/Co (∼20) values (McDonough, 2016), we set the bulk Mars composition to these values.

Abundances of moderately volatile, lithophile elements in the BSM, which can be directly converted to the bulk Mars composition after correcting for the core mass fraction, define a robust depletion trend (Figure 6). The volatility trend provides a method to determine the rest of the element abundances, except for the atmophiles, in the bulk Mars and its core (Tables 8 to 10). For the refractory siderophile and chalcophile elements (Table 1), bulk Mars abundances are set at ∼1.85 times CI abundance.

The possible conditions of Mars core formation (e.g., 10–17 GPa, 1900–2300 K, fO₂ = IW − 2 to −1) are considered to be markedly different than that for the Earth’s core (Righter and Chabot, 2011; Rai and van Westrenen, 2013). These findings (i.e., wholly lithophile character of Mn, Cr and V) are also in harmony with the relatively oxidized conditions for Mars’ core formation as compared to the Earth (e.g., Wadhwa, 2001, 2008; Herd et al., 2002).

5.2 Comparison of the core models

Although cosmochemical and geodetic insights constrain the composition and interior structure of Mars, we cannot determine a unique core model composition, given the available data. Thus, other Martian core models, distinct from that proposed in this study (Table 8), are viable. We tested three (end-member like) core compositions with (1) no H and O, (2) 10 wt% O, and (3) 1.4 wt% H, in addition to our preferred one (Table 11). There is a non-uniqueness in core mass fraction – core radius – core density, with limits being ∼15 to 26%, ∼1500 to 2000 km, and ∼5500 to 7500 kg/m³, respectively. Composition and physical properties of core models were constrained by a
combination of our BSM model, volatility trend, and geodetic properties of the planet. Importantly, as discussed in Section 5.1, the Martian volatility trend puts strong constraint on the S content in the Martian core (Figure 8).

Core models assuming H- and O-free have the highest density ($\sim 8100$ kg/m$^3$), given only S as the light element. A dense core forces down its mass fraction ($\sim 15$ wt%) and low bulk Fe/Si (1.21) and Fe/Al (13.6) ratios, as compared to most chondritic meteorites (1.0–1.8 and 15–25, respectively; Wasson and Kallemeyn [1988]). An exception, CV chondrites, have low Fe/Al ratio (13.4) due to its high abundance of refractory inclusions (i.e., high Al concentration). In contrast, however, the Fe/Si ratio in CV chondrites (1.36) is higher than that of Mars containing a core that is H- and O-free. Thus, to explain such a Mars’ bulk composition requires a chemical fractionation process not recorded in chondritic meteorites. Similar arguments can be applied for the H-bearing, O-free core model.

Physical properties of an O-rich, H-free core model is similar to that of the O- and H-bearing model proposed as the best core model. However, solubility of oxygen in liquid iron does not support such a high O concentration in the Martian core (Rubie et al., 2004; Tsuno et al., 2011).

6 Heat production in Mars

Our model predicts that the present-day heat-producing elements (HPE: K, Th and U) in the BSM produce 2.5 TW heat (Table 3). In Mars, K is a dominant radiogenic heat source during its first 3.5 Gyr history (Figure 10). Based on the BSM abundances of $^{40}$K, $^{232}$Th, $^{235}$U, $^{238}$U and $^{87}$Rb, we estimate the Martian antineutrino (or areoneutrino) luminosity is $7.7 \times 10^{24}$ $\bar{\nu}_e$/s.

7 Conclusions and implications for future works

Compositional modeling of Mars reveals that the bulk silicate Mars is enriched in refractory lithophile elements at 2.26 times higher than that in CI carbonaceous chondrites. Moderately volatile elements are systematically depleted in Mars as a function of their volatility compared to the chondritic composition, but less so than in the Earth. The Martian core contains S, O and H as light elements, which is consistent with the volatility trend and occurrence of ringwoodite at the Martian core-mantle boundary.

The chemical compositions of solar system bodies record accretion of solar nebular materials, core-mantle and mantle-crust differentiation and subsequent surface processes. The physicochemical similarities and differences between Mars and Earth provide insights into the origin and evolution of terrestrial planets, which are discussed in a companion paper (Yoshizaki and McDonough submitted).
To constrain further the interior structure and composition of Mars direct evidence from the planet is needed. The best constraints would be provided by seismic determination of the depth of Martian core-mantle boundary, which would immediately define the core’s mass fraction and density. NASA’s ongoing Interior Exploration using Investigations, Geodesy and Heat Transport (InSight) mission will provide significant constraints on the Martian core composition (Smrekar et al., 2019). In turn, our model for composition and interior structure of Mars can be tested by seismic and surface heat flux data from the InSight mission. In addition, rock samples which will be returned from Phobos in the Japan Aerospace Exploration Agency’s planned Martian Moons eXploration (MMX) mission (Kuramoto et al., 2018) are keys to further constrain not only the origin of the Martian moons (Murchie et al., 2014), but also composition of the Martian mantle, if Phobos has formed via giant impacts (Citron et al., 2015; Canup and Salmon, 2018; Hyodo et al., 2018).

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Author contributions

TY and WFM proposed and conceived various portions of this study and together calculated the compositional model of Mars. The manuscript was jointly written by TY and WFM and they read and approved the final manuscript.

Competing interests

The authors declare no competing interests.

Data and materials availability

Materials used in this study are provided as supplementary materials.
Figure 1: Ratios vs ratios of refractory lithophile elements in shergottites. The values are normalized to CI chondrite abundance (Table 2). Horizontal and vertical gray bands show CI ratio ± 10%. Trend lines for all shergottites and lherzolitic shergottites, shown in solid black and broken purple lines, respectively, cross CI chondritic compositions, showing that these chemical trends reflect melt-residue differentiation in the Martian silicate mantle. SHE-L, SHE-B and SHE-Ol are lherzolitic, basaltic and olivine-phyric shergottites, respectively.
Figure 2: Ratios vs abundances of refractory lithophile elements in lherzolitic shergottites. The values are normalized to CI chondrite ratios/abundance (Table 2). Horizontal gray bands show CI ratio ± 10%. The correlations among multiple element combinations indicate that refractory lithophile element abundance in the BSM is 2.26 times higher than in the CI chondrites (cf. 2.75 × CI in the BSE [McDonough and Sun 1995]).
Figure 3: MgO contents vs CI-normalized abundances of refractory lithophile elements in shergottites. Horizontal gray bands show the CI-normalized abundance of refractory lithophile elements in the BSM with 10% uncertainties. Intercepts of the correlation lines and the BSM composition suggest 31.0 ± 2.0 wt% MgO in the BSM. SHE-L, SHE-B and SHE-Ol are lherzolitic, basaltic and olivine-phyric shergottites, respectively.
Figure 4: SiO\textsubscript{2} data from shergottites and GRS survey of Martian surface. The box plots show median, minimum, maximum, first and last quartile of SiO\textsubscript{2} contents in shergottites. Outliers which are more than 1.5 times the interquartile range from the end of boxes are shown in dots. The Martian surface composition is determined by gamma-ray spectroscopy (GRS) (Boynton et al., 2007), whose error is in 1 standard deviation. Horizontal line and gray bands show BSM abundance of SiO\textsubscript{2} of 45.5 ± 1.8 wt%. SHE-L, SHE-B and SHE-OI are lherzolitic, basaltic and olivine-phyric shergottites, respectively.
**Figure 5:** Initial $^{87}\text{Sr}/^{86}\text{Sr}$ vs $\varepsilon^{143}\text{Nd}$ of Martian meteorites. Modified after Day et al. (2018). Compositions of individual shergottite samples and a compositional range for nakhlites and chassignites are shown. Isotopic compositions of terrestrial basalts (mid-ocean ridge basalt (MORB) and ocean island basalts (OIB) are also shown. Assuming that Martian meteorites represent isotopic composition of the Martian mantle and Rb/Sr fractionation at 4.56 Ga, the BSM abundance of Rb is estimated to be $1.6 \pm 0.8$ ppm Rb. $\varepsilon^{143}\text{Nd} = [(^{143}\text{Nd}/^{144}\text{Nd})_{\text{sample}} ÷ (^{143}\text{Nd}/^{144}\text{Nd})_{\text{chondritic}} - 1] \times 10^4$. Data are from Day et al. (2018) and references therein.
Figure 6: Abundance of lithophile (purple), siderophile (pink) and chalcophile (blue) elements in the bulk silicate Mars (Table 5) are normalized to CI chondrites (Table 2) and plotted against log of the 50% condensation temperature (K) at 10 Pa (Lodders, 2003). As shown by a purple arrow, condensation temperature of CI could be significantly lower than that proposed by Lodders (2003), which is consistent with the bulk silicate Earth composition (Clay et al., 2017) and a recent condensation calculation (Wood et al., 2019).
Figure 7: Mineralogy and physical properties of the Martian mantle. (A) Phase transitions in the Martian mantle. A red line shows an areotherm (Martian geotherm). (B) Depth versus P- and S-wave velocities and density from surface to the core-mantle boundary of Mars. Abbreviations: Ol–olivine; Wad–wadsleyite; Ring–ringwoodite; Gt–garnet; Cpx–clinopyroxene; Opx–orthopyroxene; C2/c–high-pressure clinopyroxene; Pl–plagioclase; Sp–spinel; Ca-pv–Ca-perovskite; St–stishovite.
Figure 8: Volatility trends for Mars (this study) and Earth (McDonough and Sun, 1995) constrain S contents in the metallic cores. The S-rich Martian core models (>15 wt%; e.g., Wänke and Dreibus, 1994; Taylor, 2013) are not consistent with the Martian volatility trend.
Figure 9: Interior structure of Mars. Abbreviations: Ol–olivine; Px–pyroxene; Gt–garnet; Wad–wadsleyite; Ring–ringwoodite; Brg–bridgmanite; Cpv–Ca-perovskite; Ppv–post-perovskite; Fper–ferropericlase.
Figure 10: Radiogenic heat production in the Martian mantle through time. The inset shows relative contributions of heat-producing elements as functions of time.
| 50% Tc (K) | Lithophile | Siderophile | Chalcophile |
|------------|------------|-------------|-------------|
| Refractory | 1821–1355  | Zr, Hf, Sc, Y, Gd, Tb, Dy, Ho, Er, Tm, Lu, Th, Al, U, Nd, Sm, Ti, Pr, La, Ta, Nb, Ca, Yb, Ce, Sr, Ba, Be, Eu | Re, Os, W, Ir, Mo, Ru, Pt, Rh |
| Major component | 1355–1250 | Mg, Fe, Si, Cr | Ni, Co, Fe, Pd, Cr |
| Moderately volatile | 1250-600 | Mn, Li, K, Na, Ga, Cr | P, Mn, As, Au, Cu, Ag, Sb, Zn, Te, Se, S, Cd |
| Volatile | 600–252 | Br, I, Tl | In, Tl, Hg |

| 50% Tc (K) | Atmophile |
|------------|-----------|
| Highly volatile | <252 | O, N, Xe, Kr, Ar, C, Ne, He, H |
| Highly volatile | <252 | O, N, Xe, Kr, Ar, C, Ne, He, H |

*a* Condensation temperature of Cl might be as low as Br and I ([Clay et al., 2017](#) [Wood et al., 2019](#)).

*b* Condensation temperature of In might be ∼800 K ([Righter et al., 2017](#)).
Table 2: Composition of CI chondrites adopted in this study.

| Element | Unit | Value | Reference  | Element | Unit | Value | Reference  |
|---------|------|-------|------------|---------|------|-------|------------|
| H       | %    | 1.97  | P14        | Rh      | ppm  | 0.132 | P14        |
| Li      | ppm  | 1.45  | P14        | Pd      | ppm  | 0.56  | P14        |
| Be      | ppm  | 0.0219| P14        | Ag      | ppm  | 0.201 | P14        |
| B       | ppm  | 0.775 | P14        | Cd      | ppm  | 0.674 | P14        |
| C       | ppm  | 34800 | P14        | In      | ppm  | 0.0778| P14        |
| N       | ppm  | 2950  | P14        | Sn      | ppm  | 1.63  | P14        |
| O       | %    | 45.9  | P14        | Sb      | ppm  | 0.145 | P14        |
| F       | ppm  | 58.2  | P14        | Te      | ppm  | 2.28  | P14        |
| Na      | ppm  | 4962  | P14        | I       | ppm  | 0.057 | C17        |
| Mg      | %    | 9.54  | P14        | Cs      | ppm  | 0.188 | P14        |
| Al      | %    | 0.840 | P14        | Ba      | ppm  | 2.42  | P14        |
| Si      | %    | 10.7  | P14        | La      | ppm  | 0.241 | P14        |
| P       | ppm  | 985   | P14        | Ce      | ppm  | 0.619 | P14        |
| S       | ppm  | 53500 | P14        | Pr      | ppm  | 0.0939| P14        |
| Cl      | ppm  | 115   | C17        | Nd      | ppm  | 0.474 | P14        |
| K       | ppm  | 546   | P14        | Sm      | ppm  | 0.154 | P14        |
| Ca      | %    | 0.911 | P14        | Eu      | ppm  | 0.0588| P14        |
| Sc      | ppm  | 5.81  | P14        | Gd      | ppm  | 0.207 | P14        |
| Ti      | ppm  | 447   | P14        | Tb      | ppm  | 0.0380| P14        |
| V       | ppm  | 54.6  | P14        | Dy      | ppm  | 0.256 | P14        |
| Cr      | ppm  | 2623  | P14        | Ho      | ppm  | 0.0564| P14        |
| Mn      | ppm  | 1916  | P14        | Er      | ppm  | 0.166 | P14        |
| Fe      | %    | 18.66 | P14        | Tm      | ppm  | 0.0261| P14        |
| Co      | ppm  | 513   | P14        | Yb      | ppm  | 0.169 | P14        |
| Ni      | ppm  | 10910 | P14        | Lu      | ppm  | 0.0250| P14        |
| Cu      | ppm  | 133   | P14        | Hf      | ppm  | 0.107 | P14        |
| Zn      | ppm  | 309   | P14        | Ta      | ppm  | 0.0150| P14        |
| Ga      | ppm  | 9.62  | P14        | W       | ppm  | 0.0960| P14        |
| Ge      | ppm  | 32.6  | P14        | Re      | ppm  | 0.0381| P14        |
| As      | ppm  | 1.74  | P14        | Os      | ppm  | 0.461 | P14        |
| Se      | ppm  | 20.3  | P14        | Ir      | ppm  | 0.431 | P14        |
| Br      | ppm  | 0.189 | C17        | Pt      | ppm  | 0.874 | P14        |
| Rb      | ppm  | 2.32  | P14        | Au      | ppm  | 0.175 | P14        |
| Sr      | ppm  | 7.79  | P14        | Hg      | ppm  | 0.35  | P14        |
| Y       | ppm  | 1.46  | P14        | Tl      | ppm  | 0.169 | W15        |
| Zr      | ppm  | 3.63  | P14        | Pb      | ppm  | 2.62  | P14        |
| Nb      | ppm  | 0.283 | P14        | Bi      | ppm  | 0.087 | W15        |
| Mo      | ppm  | 0.87  | W15        | Th      | ppm  | 0.0300| P14        |
| Ru      | ppm  | 0.69  | P14        | U       | ppm  | 0.00796| W18       |

*Clay et al. (2017), Day et al. (2016), Palme et al. (2014), Wang et al. (2015), Wipperfurth et al. (2018).*
Table 3: Comparison of compositional models of the bulk silicate Mars.

| wt%  | This study | WD94 | T13 | MA79 | OK92 | LF97 | S99 | KC08 |
|------|------------|------|-----|------|------|------|-----|------|
| SiO₂ | 45.5       | 44.4 | 43.7| 41.6 | 43.0 | 45.4 | 47.5| 44   |
| TiO₂ | 0.17       | 0.14 | 0.14| 0.33 | 0.24 | 0.14 | 0.1 | -    |
| Al₂O₃| 3.59       | 3.02 | 3.04| 6.39 | 3.48 | 2.89 | 2.5 | 2.5  |
| MnO  | 0.37       | 0.46 | 0.44| 0.15 | 0.22 | 0.37 | 0.4 | -    |
| FeO  | 14.7       | 17.9 | 18.1| 15.9 | 15.1 | 17.2 | 17.7| 17   |
| MgO  | 31.0       | 30.2 | 30.5| 29.8 | 34.3 | 29.7 | 27.3| 33   |
| CaO  | 2.88       | 2.45 | 2.43| 5.16 | 2.81 | 2.36 | 2.0 | 2.2  |
| Na₂O | 0.59       | 0.5  | 0.53| 0.10 | 0.46 | 0.98 | 1.2 | -    |
| K₂O  | 0.043      | 0.037| 0.037|0.009 | - | 0.11 | - | -   |
| P₂O₅ | 0.17       | 0.16 | 0.19| -    | -   | 0.18 | - | -    |
| NiO  | 0.046      | -    | 0.03| -    | -   | -    | - | -    |
| Cr₂O₃| 0.88       | 0.76 | 0.73| 0.65 | 0.40 | 0.68 | 0.7 | -    |
| K (ppm)| 360       | 305 | 309| 76.5 | - | 920 | - | -   |
| Th (ppb)| 68        | 56  | 58 | 125 | - | 55  | - | -   |
| U (ppb)| 18        | 16  | 16 | 35  | - | 16  | - | -   |
| Total | 99.9      | 100 | 99.8| 100  | 100 | 100 | 100| 98.7 |
| Mg#  | 0.79       | 0.75 | 0.75| 0.77 | 0.75 | 0.76 | 0.72| 0.77 |
| Mg/Si | 0.88      | 0.88 | 0.90| 0.92 | 1.03 | 0.84 | 0.74| 0.97 |
| Al/Si | 0.09      | 0.08 | 0.08| 0.17 | 0.09 | 0.07 | 0.06| 0.06 |
| RLE/C  | 2.3       | 2.1  | 2.1| 4.4  | 2.4 | 2.0  | 1.7 | 1.9  |
| Fe/Si | 0.54      | 0.67 | 0.69| 0.63 | 0.58 | 0.63 | 0.62| 0.64 |
| Fe/Al | 6.0       | 8.7  | 8.7| 3.6  | 6.4 | 8.8  | 10.4| 10.0 |
| H₉BSM| 2.5       | 2.0  | 2.0| 3.4  | -  | 3.0  | - | -    |
| H₉BSM (pW/kg) | 4.8 | 4.1  | 4.2| 7.0  | -  | 6.2  | - | -    |

a Wänke and Dreibus (1994).  b Taylor (2013).  c Morgan and Anders (1979).  d Ohtani and Kamaya (1992).  e Lodders and Fegley Jr (1997).  f Sanloup et al. (1999) (EH45:H55 model).  g Khan and Connolly (2008).  h An average enrichment factor of major refractory lithophile elements (Ca, Al and Ti) compared to CI abundance (Table 2).  i Heat production in the bulk silicate Mars (BSM: mantle + crust).
Table 4: Major element composition of the bulk silicate Mars (BSM). See text for the details of methods used to determine the BSM abundance of elements.

| Element | wt%  | 1sd | rsd% | Method |
|---------|------|-----|------|--------|
| SiO$_2$ | 45.5 | 1.8 | 4    | Mean SHE |
| TiO$_2$ | 0.17 | 0.02| 10   | RLE    |
| Al$_2$O$_3$ | 3.59 | 0.36| 10   | RLE    |
| MnO    | 0.37 | 0.07| 18   | FeO/MnO in SNC |
| FeO    | 14.7 | 1.0 | 7    | Mg# = 0.79 ± 0.02$^a$ |
| MgO    | 31.0 | 2.0 | 6    | vs RLE in SHE |
| CaO    | 2.88 | 0.29| 10   | RLE    |
| Na$_2$O | 0.59 | 0.13| 22   | Na/Al in SHE |
| K$_2$O | 0.043| 0.005| 11 | K/Th (GRS$^c$) |
| P$_2$O$_5$ | 0.17 | 0.05| 28   | P/Y in SHE |
| NiO    | 0.046| 0.01| 26   | Ni/Mg in SHE-L and SHE-Ol |
| Cr$_2$O$_3$ | 0.88 | 0.15| 17   | vs Al in SHE-L and SHE-Ol |

Total 99.9

Mg# 0.79
Mg/Si 0.88
Al/Si 0.09
RLE/CI 2.26
Fe/Si 0.54
Fe/Al 6.0

$^a$ GRS–gamma-ray spectroscopy; RLE–refractory lithophile elements; SHE-L–lherzolitic shergottites; SHE-Ol–olivine shergottites; SNC–shergottite, nakhlite and chassignite.

$^b$ Borg and Draper (2003); Agee and Draper (2004); Draper et al. (2005); Minitti et al. (2006); White et al. (2006); McCoy et al. (2016).

$^c$ Taylor et al. (2006a,b).
Table 5: Composition of the bulk silicate Mars. Concentrations are in ppm (µg/g), otherwise noted. See text for the details of methods used to determine the BSM abundances of elements.

| Element | BSM | rsd% | Method | Element | BSM | rsd% | Method |
|---------|-----|------|--------|---------|-----|------|--------|
| H       | 16  | U    | See text | Rh      | 0.0021 | 4    | Mean SNC |
| Li      | 1.8 | 22   | vs RLE in SNC | Pd      | 0.0074 | 70   | vs MgO |
| Be      | 0.05| 10   | RLE     | Ag      | 0.02  | U    | Sh/Na in SHE |
| B       | 0.84| 63   | R/Ca in SHE-L and SHE-Ol | Cd      | 0.020 | 50   | Cd/Yb in SHE |
| C       | 32  | U    | Chondritic C/H | In      | 0.010 | 50   | In/Y in SHE |
| N       | 1.6 | U    | Chondritic N/H | Sn      | 0.28  | 40   | Sn/Sn in SHE |
| O (%)   | 43.2| 8    | Major oxide stoichiometry, E<sub>87</sub>Sr<sup>86</sup>Sr = 0.7 | Sh      | 0.02  | 50   | Sh/Ps in SHE |
| F       | 28  | 35   | Cl/F in SNC | Te      | 0.0005 | 50   | Cu/Te and Se/Te in SHE |
| Na      | 4380| 22   | Na/Al in SHE | I       | 0.009 | U    | Volatility trend |
| Mg (%)  | 18.7| 6    | vs RLE in SHE | Cs      | 0.07  | 41   | Cs/La in SHE |
| Al (%)  | 1.90| 10   | RLE     | Ba      | 5.47  | 10   | RLE |
| Si (%)  | 21.3| 4    | Mean SHE | La      | 0.546 | 10   | RLE |
| P       | 740 | 28   | PY in SHE | Ce      | 1.40  | 10   | RLE |
| S       | 360 | 33   | S/Cu and S/Se in SHE | Pr      | 0.212 | 10   | RLE |
| Cl      | 28  | 35   | Cl/la in SHE | Nd      | 1.07  | 10   | RLE |
| K       | 360 | 11   | K/Ti (GRS) | Sm      | 0.347 | 10   | RLE |
| Ca (%)  | 2.06| 10   | RLE     | Eu      | 0.133 | 10   | RLE |
| Sc      | 13.1| 10   | RLE     | Gd      | 0.468 | 10   | RLE |
| Ti      | 1010| 10   | RLE     | Tb      | 0.0858 | 10   | RLE |
| V       | 123 | 10   | RLE     | Dy      | 0.578 | 10   | RLE |
| Cr      | 6000| 17   | vs Al in SHE-L and SHE-Ol | Ho      | 0.128 | 10   | RLE |
| Mn      | 2880| 18   | FeO/MnO in SNC | Er      | 0.374 | 10   | RLE |
| Fe (%)  | 11.4| 7    | Mg# = 0.79 ± 0.1 | Tm      | 0.0950 | 10   | RLE |
| Co      | 96  | 46   | Co/Fe in SHE-L and SHE-Ol | Yb      | 0.381 | 10   | RLE |
| Ni      | 360 | 26   | Ni/Fe in SHE-L and SHE-Ol | Lu      | 0.0566 | 10   | RLE |
| Cu      | 2.6 | 23   | Cu/Ti in SHE | Hf      | 0.241 | 10   | RLE |
| Zn      | 45  | 33   | vs La in SHE | Ta      | 0.0359 | 14   | RLE |
| Ga      | 8.7 | 22   | Ga/Al in SHE | La      | 0.069 | 54   | 13/Si in SHE |
| Ge      | 0.6 | 67   | vs MgO in SHE-L and SHE-Ol | Re      | 0.0004 | 75   | vs MgO |
| As      | 0.03 | 76 | As/Al in SNC | Oe      | 0.004 | 50   | vs MgO |
| Se      | 0.10 | 27 | Cu/Se in SHE | Ir      | 0.0037 | 60   | vs MgO |
| Br      | 0.12 | 72 | Cl/Br in SHE | Pt      | 0.0096 | 80   | vs MgO |
| Rb      | 1.2 | 36 | 87Sr/86Sr = 0.7 | Au      | 0.002 | 80   | vs MgO |
| Sr      | 17.6| 10   | RLE     | Hg      | 0.007 | U    | Hg/Se in NAK and the BSE |
| Y       | 3.30| 10   | RLE     | Tl      | 0.004 | 50   | Tl/Sn in SHE |
| Zr      | 8.20| 10   | RLE     | Pb      | 0.255 | 18   | U-Pb isotope systematics in SNC |
| Nb      | 0.640| 10   | RLE     | Bi      | 0.002 | 60   | Bi/Th in SHE |
| Mo      | 0.5 | 80 | Mo/Cr in SHE | Th      | 0.0678 | 10   | RLE |
| Ru      | 0.0062 | 60   | vs MgO | U      | 0.0180 | 10   | RLE |

<sup>a</sup>U–uncertain. <sup>b</sup>McCubbin et al. 2016a,b; Filiberto et al. 2019. <sup>c</sup>FeO–bulk silicate Earth. GRS–gamma-ray spectroscopy; NAK–nakhlite; SNC–shergottite, herzolitic anorthosites; SHE-Ol–olivine shergottites. <sup>d</sup>Schmidt et al. 2013; Medard et al. 2014. <sup>e</sup>Medard et al. 2014; Wang and Becker 2017. <sup>f</sup>Filiberto et al. 2016; this study. <sup>g</sup>Tait and Day 2018. <sup>h</sup>Yang et al. 2015. <sup>i</sup>Borg and Draper 2003; Agee and Draper 2004; Draper et al. 2005; Minuti et al. 2006; White et al. 2006; McCoy et al. 2016. <sup>j</sup>Righter and Chabot 2011. <sup>k</sup>Tait and Day 2018. <sup>l</sup>µ = 3.6, 208Pb/204Pb = 33, 207Pb/204Pb = 13, 206Pb/204Pb = 14. 

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Table 6: Comparison of compositional models of the Martian core.

|      | This study | WD94 | T13 | MA79 | OK92 | LF97 | S99 | KC08 |
|------|------------|------|-----|------|------|------|-----|------|
| Fe   | 79.5       | 77.8 | 78.6| 88.1 | 78.4 | 81.1 | 76.6| 75–78 |
| Ni   | 7.4        | 7.6  | -   | 8.0  | 7.6  | 7.7  | 7.2 | -    |
| Co   | 0.33       | 0.36 | -   | 0.37 | -    | 0.4  | -   | -    |
| S    | 6.6        | 14.24| 21.4| 3.5  | 14.0 | 10.6 | 16.2| 22–25 |
| P    | 0.33       | -    | -   | -    | -    | 0.2  | -   | -    |
| O    | 5          | -    | -   | -    | -    | -    | -   | -    |
| H    | 0.8        | -    | -   | -    | -    | -    | -   | -    |
| Total| 99.9       | 100  | 100 | 100  | 100  | 100  | 100 | -    |

Notes:

- a Wänke and Dreibus (1994).
- b Taylor (2013).
- c Morgan and Anders (1979).
- d Ohtani and Kamaya (1992).
- e Lodders and Fegley Jr (1997).
- f Sanloup et al. (1999) (EH45:H55 model).
- g Khan and Connolly (2008).

h Sum of Fe and Ni.
Table 7: Physical properties of Mars and Earth. Modeled values are in a normal font and reference values are in italic.

| Observation | Unit | Crust | Mantle | Core | Bulk planet | Reference value |
|-------------|------|-------|--------|------|-------------|-----------------|
| **Mars**    |      |       |        |      |             |                 |
| Mass        | kg   | $2.56 \times 10^{22}$ | $5.01 \times 10^{23}$ | $1.17 \times 10^{23}$ | $6.418 \times 10^{23}$ | $6.41856(8) \times 10^{23}$ |
| Mean density| kg/m$^3$ | 3010 | 3640 | 7060 | 3935 | $3.935(1)$ |
| Moment of inertia| – | 7% | 89% | 4% | 0.3638 | 0.3639(1) |
| Heat production (K, Th, U) | TW | 1.3 | 1.3 | 0 | 2.5 | 2.9(1) |
| Heat production (K, Th, U) | pW/kg | $4.6^{\text{a}}$ | $2.5^{\text{c}}$ | 0 | $3.9^{\text{f}}$ | – |
| **Earth**   |      |       |        |      |             |                 |
| Mass        | kg   | $3.12 \times 10^{22}$ | $4.00 \times 10^{24}$ | $1.94 \times 10^{24}$ | $5.97 \times 10^{24}$ | $5.97218(60) \times 10^{24}$ |
| Mean density| kg/m$^3$ | 2800 | 4400 | 11870 | 5510 | 5510 |
| Moment of inertia| – | 1% | 88% | 11% | 0.3308 | 0.3308 |
| Heat production (K, Th, U) | TW | $7.5^{\text{a}}$ | $12.6^{\text{c}}$ | 0 | $19.9^{\text{i}}$ | 4(1) |
| Heat production (K, Th, U) | pW/kg | 232 | 3.1 | 0 | 3.3 | – |

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$a$ Rivoldini et al. (2011). $b$ Konopliv et al. (2016). $c$ Surface heat loss estimated based on an average surface heat flux of 20 ± 1 mW/m$^2$ (Parro et al., 2017; Samuel et al., 2019). $d$ Abundances of heat-producing elements (HPE) in the crust: 3740 ppm K, 700 ppb Th and 180 ppb U (Taylor and McLennan, 2009). $e$ HPE abundance in the mantle is calculated from mass-balance considerations: 190 ppm K, 36 ppb Th and 10 ppb U. $f$ BSM abundances of HPE are 360 ppm K, 68 ppb Th and 18 ppb U (this study). $g$ Chambat et al. (2010). $h$ The preliminary Earth model (Dziewonski and Anderson, 1981). $i$ Huang et al. (2013). $j$ Jaupart et al. (2015).
Table 8: Composition of the Martian core. Concentrations are in ppm (µg/g), otherwise noted.

| Element | Martian core | Element | Martian core |
|---------|--------------|---------|--------------|
| Fe (%)  | 79.5         | Os      | 5            |
| Ni (%)  | 7.4          | Pd      | 5            |
| O (%)   | 5.0          | Ir      | 5            |
| S (%)   | 6.6          | Te      | 3            |
| H (%)   | 0.8          | Pb      | 3.1          |
| Co      | 3300         | Rh      | 1.3          |
| P       | 3300         | Sn      | 1.3          |
| Cu      | 560          | W       | 0.7          |
| Zn      | 290          | Ag      | 0.7          |
| Ge      | 90           | Au      | 0.7          |
| Se      | 30           | Cd      | 0.7          |
| Pt      | 9            | Re      | 0.4          |
| As      | 8            | Sb      | 0.4          |
| Mo      | 7            | Bi      | 0.1          |
| Ru      | 7            | Tl      | 0.1          |
Table 9: Composition of the bulk Mars. Concentrations are in ppm (µg/g), otherwise noted.

| Element | Bulk Mars | Element | Bulk Mars |
|---------|-----------|---------|-----------|
| H       | 1500      | Rh      | 0.24      |
| Li      | 1.5       | Pd      | 0.84      |
| Be      | 0.041     | Ag      | 0.14      |
| B       | 0.69      | Cd      | 0.15      |
| C       | 26        | In      | 0.01      |
| N       | 1.4       | Sn      | 0.43      |
| O (%)   | 36.3      | Sb      | 0.1       |
| F       | 23        | Te      | 0.6       |
| Na      | 3600      | I       | 0.01      |
| Mg (%)  | 15.3      | Cs      | 0.06      |
| Al (%)  | 1.56      | Ba      | 4.5       |
| Si (%)  | 17.4      | La      | 0.45      |
| P       | 1200      | Ce      | 1.1       |
| S       | 12100     | Pr      | 0.17      |
| Cl      | 23        | Nd      | 0.88      |
| K       | 300       | Sm      | 0.28      |
| Ca (%)  | 1.69      | Eu      | 0.11      |
| Sc      | 10.8      | Gd      | 0.38      |
| Ti      | 830       | Tb      | 0.070     |
| V       | 100       | Dy      | 0.47      |
| Cr      | 4900      | Ho      | 0.10      |
| Mn      | 2400      | Er      | 0.31      |
| Fe (%)  | 23.7      | Tm      | 0.048     |
| Co      | 680       | Yb      | 0.31      |
| Ni      | 13600     | Lu      | 0.046     |
| Cu      | 100       | Hf      | 0.20      |
| Zn      | 89        | Ta      | 0.028     |
| Ga      | 7.2       | W       | 0.18      |
| Ge      | 16        | Re      | 0.07      |
| As      | 1.4       | Os      | 0.9       |
| Se      | 5.2       | Ir      | 0.9       |
| Br      | 0.1       | Pt      | 1.7       |
| Rb      | 0.96      | Au      | 0.1       |
| Sr      | 14        | Hg      | 0.006     |
| Y       | 2.7       | Tl      | 0.02      |
| Zr      | 6.7       | Pb      | 0.8       |
| Nb      | 0.52      | Bi      | 0.03      |
| Mo      | 1.6       | Th      | 0.056     |
| Ru      | 1.3       | U       | 0.015     |
Table 10: Compositional models for the bulk Mars, bulk silicate Mars (BSM) and core and atomic proportions for major elements.

|         | wt% | Bulk | BSM | Core | atomic% | Bulk | BSM | Core |
|---------|-----|------|-----|------|---------|------|-----|------|
| O       | 36.3| 43.2 | 5.0 | O    | 53      | 59   | 11  |
| Mg      | 15.3| 18.7 | 0   | Mg   | 15      | 17   | 0   |
| Si      | 17.4| 21.3 | 0   | Si   | 15      | 17   | 0   |
| Fe      | 23.7| 11.4 | 79.5| Fe   | 10      | 4.5  | 50  |
| Al      | 1.56| 1.90 | 0   | Al   | 1.4     | 1.5  | 0   |
| S       | 1.2 | 0.04 | 6.6 | S    | 0.9     | 0.02 | 7.1 |
| Ca      | 1.69| 2.06 | 0   | Ca   | 1.0     | 1.1  | 0   |
| Ni      | 1.4 | 0.04 | 7.4 | Ni   | 0.5     | 0.01 | 4.4 |
| H       | 0.1 | 0.00 | 0.8 | H    | 3.4     | 0.03 | 28  |
Table 11: Comparison of chemical and physical properties of Martian core models.

| Chemical composition | H₂O-bearing | H₂O-free | H-free | O-free | Reference value |
|----------------------|-------------|----------|--------|--------|-----------------|
| Core wt%              |             |          |        |        |                 |
| Fe                   | 79.5        | 81.6     | 76.0   | 82.4   |                 |
| Ni                   | 7.4         | 8.2      | 7.0    | 7.8    |                 |
| Co                   | 0.33        | 0.37     | 0.32   | 0.35   |                 |
| S                    | 6.56        | 8.17     | 6.14   | 7.04   |                 |
| P                    | 0.33        | 0.41     | 0.31   | 0.36   |                 |
| O                    | 5.0         | 0.0      | 10.0   | 0      |                 |
| H                    | 0.8         | 0.0      | 0.0    | 1.4    |                 |
| Total                | 99.9        | 98.8     | 99.8   | 99.3   |                 |

| Bulk Mars             |             |          |        |        |                 |
| Fe/Si                | 1.36        | 1.21     | 1.37   | 1.33   | 1.0–1.4          |
| Fe/Al                | 15.2        | 13.6     | 15.4   | 14.9   | 15.5–24.3       |

| Modeled physical properties |             |          |        |        |                 |
|------------------------------|-------------|----------|--------|--------|-----------------|
| $M'_\text{core}$ (mass fr.) | 18%         | 15%      | 19%    | 17%    |                 |
| $\rho_{\text{core}}$ (kg/m$^3$) | 7.060      | 8.120    | 6.940  | 7.320  |                 |
| $r_{\text{core}}$ (km)       | 1.580       | 1.400    | 1.610  | 1.520  |                 |
| $P_{\text{CMB}}$ (GPa)       | 22          | 24       | 21     | 22     |                 |
| $M_{\text{bulk}}$ (kg)       | $6.420 \times 10^{23}$ | $6.414 \times 10^{23}$ | $6.427 \times 10^{23}$ | $6.423 \times 10^{23}$ | $6.419 \times 10^{23}$ |
| $\rho_{\text{bulk}}$ (kg/m$^3$) | 3.935      | 3.932    | 3.939  | 3.947  | 3.935           |
| MOI                        | 0.3638      | 0.3644   | 0.3632 | 0.3627 | 0.363           |

$^a$ Compositional range of ordinary chondrites [Wasson and Kallemeyn, 1988].
$^b$ $M'$–mass fraction; $\rho$–density; $r_{\text{core}}$–radius of the core; $P_{\text{CMB}}$–pressure at the core–mantle boundary; $M$–mass.
$^c$ Rivoldini et al. (2011).
$^d$ Konopliv et al. (2016).
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52
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