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## The Temperature and Composition of the Mantle Sources of Martian Basalts

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### Key Points:

- Basalts that sampled discrete mantle regions throughout Mars's history provide information about the mantle composition and temperature
- The mantle potential temperature of primitive basalts appears constant (1400–1500°C), yet is likely not representative of the average mantle
- Incompatible element concentrations in the mantle vary due to magma ocean crystallization, partial melting, and metasomatism

### Supporting Information:

Supporting Information may be found in the online version of this article.

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**Abstract** The composition of basaltic melts in equilibrium with the mantle can be determined for several Martian meteorites and in-situ rover analyses. We use the melting model MAGMARS to reproduce these primary melts and estimate the bulk composition and temperature of the mantle regions from which they originated. We find that most mantle sources are depleted in CaO and Al<sub>2</sub>O<sub>3</sub> relative to models of the bulk silicate Mars and likely represent melting residues or magma ocean cumulates. The concentrations of Na<sub>2</sub>O, K<sub>2</sub>O, P<sub>2</sub>O<sub>5</sub>, and TiO<sub>2</sub> are variable and often less depleted, pointing to the re-fertilization of the sources by fluids and low-degree melts, or the incorporation of residual trapped melts during the crystallization of the magma ocean. The mantle potential temperatures of the sources are 1400–1500°C, regardless of the time at which they melted and within the range of the most recent predictions from thermochemical evolution models.

**Plain Language Summary** Martian meteorites and rocks analyzed by rovers are witnesses of magmatic processes on Mars. Some of the basaltic specimens among them have been classified as “primitive” as they are closely related to the melts that form in the mantle and feature high MgO/FeO. They record important properties of the mantle of Mars. We use the mantle melting model MAGMARS to constrain the temperature and composition of the mantle source regions from which primitive basalts originated. We find that the mantle compositions were low in CaO and Al<sub>2</sub>O<sub>3</sub>, either because they melted on several occasions, or because these components were locked in deeper layers of the mantle when it solidified from the bottom up (early magma ocean). Several mantle sources are comparatively rich in Na<sub>2</sub>O, K<sub>2</sub>O, P<sub>2</sub>O<sub>5</sub>, and TiO<sub>2</sub>. These components were either subsequently added to the mantle sources by fluids and low-degree melts or can be explained by the trapping of melts during the progressive crystallization of the magma ocean. The temperature of the mantle sources projected to surface conditions for easier comparison (potential temperature) was 1400–1500°C, regardless of the time at which they melted, and is within the range of recent predictions from planetary-scale models of interior dynamics.

## 1. Introduction

Our knowledge of the thermal state, composition, and structure of the Martian mantle is derived from a diverse and continuously expanding array of geophysical and geochemical constraints. Early measurements of the moment of inertia factor, soil compositions at the Viking landing sites, and the definitive recognition that shergottites, nakhlites, and chassignites are from Mars (Baird et al., 1976; Bogard & Johnson, 1983; Johnston & Toksöz, 1977), unequivocally pointed to a FeO-rich mantle ( $\text{Mg}/(\text{Fe} + \text{Mg}) \times 100$  in moles or  $\text{Mg}\# = 75\text{--}81$ ) compared to Earth (90). Model compositions of the “primitive mantle” were rapidly put forth (e.g., Dreibus & Wänke, 1985) and allowed to create simple models of the Martian interior structure (Bertka & Fei, 1997; Elkins-Tanton et al., 2003; Longhi et al., 1992). Additional analyses of crustal rocks by subsequent orbiting probes and rovers, the discovery of new Martian meteorites (Agee et al., 2013; Humayun et al., 2013), geodetic and seismic data from the recent InSight mission (e.g., Huang et al., 2022; Khan et al., 2021), and geodynamic modeling (e.g., Plesa et al., 2022), are now allowing to draw ever improving representations of the interior structure of Mars and its evolution through time.

Currently available compositions of the Martian mantle (e.g., Dreibus & Wänke, 1985; Khan et al., 2022; Lodders & Fegley, 1997; Yoshizaki & McDonough, 2020, abbreviated as DW85, K22, LF97, and YM20 hereinafter) represent average and idealized primitive compositions that are useful to derive average characteristics (density, solidus temperature, seismic wave velocity, etc.). However, the study of Martian meteorites has shown

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that the mantle is highly heterogeneous—both in terms of isotopic composition and Mg#—as a consequence of the crystallization of the Martian Magma Ocean (MMO) (e.g., Borg et al., 1997; Bouvier et al., 2018; Debaille et al., 2008; Humayun et al., 2013; Kruijjer et al., 2017; Nyquist et al., 2016; Udry et al., 2020). The major-element composition of the mantle reservoirs formed during this early differentiation event remains poorly constrained and model-dependent (e.g., Borg & Draper, 2003; Elkins-Tanton et al., 2005).

To derive more detailed models of the interior structure of Mars, independent constraints on the composition and temperature of discrete regions of the Martian mantle are desirable. A subset of Martian basalts, characterized by varied crystallization ages and high Mg# (Table 1) have been suggested to represent primitive basalts in near-equilibrium with their mantle sources and have been used to determine the  $P$ – $T$  conditions of their mantle source through experiments (Filiberto, Dasgupta, et al., 2010; Filiberto et al., 2008; Filiberto, Musselwhite, et al., 2010; Monders et al., 2007; Musselwhite et al., 2006) or modeling (Balta & McSween, 2013a; Baratoux et al., 2011; Filiberto, 2017; Filiberto & Dasgupta, 2011, 2015; Lee et al., 2009). The major element composition of these basalts cannot be produced by melting the primitive mantle and are instead expected to derive from mantle sources of diverse compositions (e.g., Schmidt & McCoy, 2010; Collinet et al., 2015; Figure 1).

Here, we use MAGMARS, a new model developed to simulate melting in the Martian mantle (Collinet et al., 2021), to re-evaluate the melting conditions and the thermal state of the mantle sources of primitive Martian basalts, which crystallized at different times and therefore represent snapshots of Mars' thermochemical evolution. In addition, MAGMARS allows us to estimate for the first time the major-element composition of these local mantle sources. We find that the  $P$ – $T$  melting conditions appear to have remained relatively stable through time and that mantle sources display variable  $\text{CaO}/\text{Al}_2\text{O}_3$ , low overall abundances of incompatible elements but enrichment of alkalis, P and Ti relative to Ca and Al. We discuss the implications of these findings for the early differentiation of Mars and its long-lived magmatism.

## 2. Selected Compositions of Primitive Martian Basalts

While the majority of mantle melts were modified by igneous differentiation as they ascended through the crust (Farley et al., 2022; Ostwald et al., 2022; Payré et al., 2020; Udry et al., 2018; Wiens et al., 2022), a limited number of Martian basalts bear witness to the composition and temperature of the mantle at the time of their formation (i.e., primitive basalts). To identify primitive basalts, we first make the assumption that the average Martian mantle contains olivine  $\text{Mg}\# \geq 77$  (Table 1, Table S1, and Figure 1), and would produce primary melts with a  $\text{Mg}\# \geq 54$  ( $K_{D, \text{Fe-Mg}}^{\text{oliv-liq}}$  of 0.35; Filiberto & Dasgupta, 2011; Matzen et al., 2022). A mantle of  $\text{Mg}\# 77$  is intermediate between the most commonly accepted primitive mantle compositions (Dreibus & Wänke, 1985; Yoshizaki & McDonough, 2020). Here, we only consider Martian basaltic compositions with a  $\text{Mg}\# \geq 48$  (Table 1), which could derive from primary mantle melts of  $\text{Mg}\# \geq 54$  following a maximum of 10 wt.% of olivine fractionation.

The *Spirit* rover analyzed numerous basalts with  $\text{Mg}\# 48$ – $55$  at Gusev crater (McSween, Wyatt, et al., 2006; Ming et al., 2008; Squyres et al., 2007) that could represent primitive basalts (Filiberto, Dasgupta, et al., 2010; Monders et al., 2007; Schmidt & McCoy, 2010). Among these, the Adirondack-class basalts are poor in  $\text{K}_2\text{O}$  and could derive from a residual mantle depleted in incompatible elements by prior melting events (Collinet et al., 2021; Schmidt & McCoy, 2010) while most of the basalts analyzed in the vicinity of the Columbia Hills are more enriched in alkali elements and poorer in CaO (Figure 1). The ancient regolith breccia Northwest Africa (NWA) 7034/7475/7533 (Bouvier et al., 2018; Cassata et al., 2018; Humayun et al., 2013; Nyquist et al., 2016) is also characterized by a high  $\text{Mg}\#$  (54; Wittmann et al., 2015) and, despite its complex history, could approach the composition of a mantle melt based on trace (Humayun et al., 2013) and major element compositions (Collinet et al., 2015). We also test whether two individual clasts, representative of the main families of basaltic clasts (Santos et al., 2015; Wittmann et al., 2015), could be primitive basalts later remelted by impacts: a vitrophyre (Udry et al., 2014) and an alkali-rich microbasalt known as “Clast VI” (Humayun et al., 2013).

Recent geophysical constraints suggest that large portions of the mantle could be more Mg-rich ( $\text{Mg}\# = 81$ ; Khan et al., 2022) than previously assumed (e.g., Dreibus & Wänke, 1985; Yoshizaki & McDonough, 2020), as also evidenced by the study of Martian meteorites. The most primitive depleted shergottite (Yamato (Y) 980459, nearly identical to NWA 5789; Greshake et al., 2004; Gross et al., 2011) and the most primitive enriched shergottite (Larkman Nunatak (LAR) 06319, nearly identical to NWA 1068; Barrat et al., 2002; Peslier et al., 2010) have  $\text{Mg}\#$  of 66 and 58, respectively. Y 980459 contains olivine  $\text{Mg}\# 85$ – $86$  and is thought to represent a primary melt

**Table 1**  
*Martian Primitive Basalts, Fraction of Olivine Addition Required to Reach Mantle-Melt Equilibrium, and Associated Inferred Mantle Mg#*

		Age (Ga)	Oliv (wt.%)	Mg#
NWA 7034	Vitrophyre [1]	4.49 [2]	+10/+26	77/81
NWA 7533	Clast VI [3]	4.49 [2]	+9/+24	77/81
NWA 7475	Bulk [4]	4.49 [2]	0	77
Adirondack-class basalts [5]		3.7 [6]	+3/+17	77/81
Columbia Hills	Humboldt Peak [7]	3.7 [6]	+7/+20	77/81
	Fastball [8]	3.7 [6]	0/+13	77/81
	Stars, etc. [8]	3.7 [6]	+5/+17	77/81
	Ace [8]	3.7 [6]	+9/+29	77/81
	Irvine [7]	3.7 [6]	+8/+25	77/81
Chassignite	NWA 2737 [9]	1.3 [10]	+9	79
Depleted shergottite	Y 980459 [11]	0.47 [12]	0/+7	85/86
Enriched shergottite	LAR 06319 [13]	0.19 [14]	0/+5	80/81

*Note.* [1] Udry et al. (2014), [2] Costa et al. (2020), [3] Humayun et al. (2013), [4] Wittmann et al. (2015), [5] McSween, Wyatt, et al. (2006), [6] Greeley et al. (2005), [7] Ming et al. (2008), [8] Squyres et al. (2007), [9] He et al. (2013), [10] Udry and Day (2018), [11] average of Misawa (2004), Shirai and Ebihara (2004), and Greshake et al. (2004), [12] Shih et al. (2005), [13] Basu Sarbadhikari et al. (2009), and [14] Shafer et al. (2010).

composition (e.g., Matzen et al., 2022; Musselwhite et al., 2006). The olivine megacrysts in LAR 06319 and NWA 1068 have Mg#  $\leq 77$  (Basu Sarbadhikari et al., 2009) but were initially more magnesian (Mg# 80) and were modified by Fe–Mg diffusion (Balta et al., 2013; Collinet et al., 2017). NWA 2737 is a chassignite (Mg# 79) with olivine-hosted melt inclusions. Its primary melt is taken as the reconstructed composition of the parental trapped liquid (He et al., 2013). Given the multitude of evidence of Mg-rich mantle reservoirs, we also calculated alternative primary melt compositions for the Gusev basalts and NWA 7034/7475/7533 bulk rock and basaltic clasts that would be in equilibrium with a mantle of Mg# 81. The basaltic compositions in equilibrium with a mantle of Mg# 77–86, following addition of 0%–29% of olivine (Table 1 and Table S1) are collectively referred to as “target compositions” thereafter.

### 3. Methods

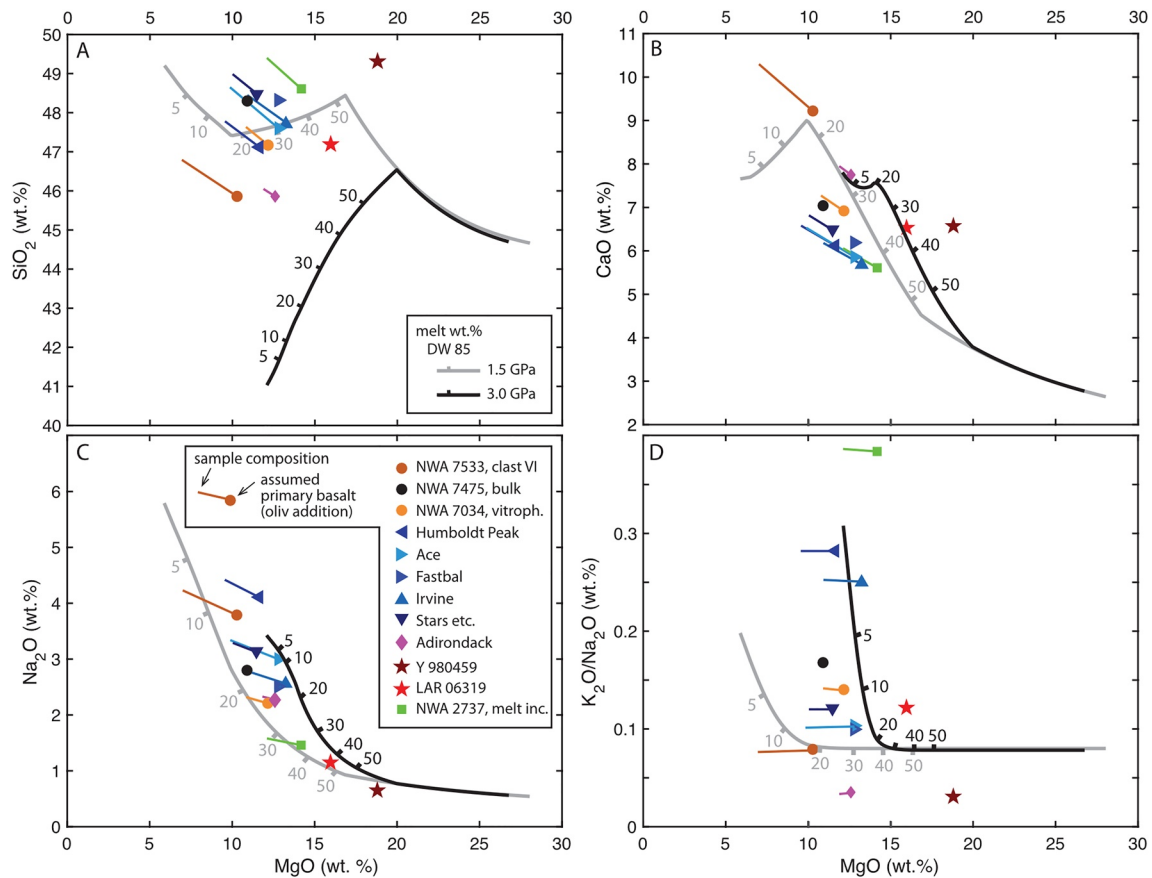
To constrain the major element composition and temperature of the mantle sources of the target compositions (Table 1 and Table S1), we start by simulating the melting of one of the primitive mantle compositions (DW85, K22, and YM20) using MAGMARS in isobaric mode (e.g., Figure 1). Second, we adjust the mantle composition incrementally until the melt produced is identical to the target composition (i.e., when the concentrations of all major and incompatible elements are within 1 wt.% relative). The first mantle composition identified that fulfill these criteria, for any given input  $T_p$  and pressure of melting, is called “source 1” (Table S2) and can deviate substantially from the primitive mantle. In a third step, we mathematically remove a fraction

(33–50 wt.%) of the melt produced from source 1 and repeat the same procedure to identify increasingly refractory mantle compositions (sources 2 and 3) that can still produce melts identical to the target compositions. Finally, we attempt to reproduce the sources constrained in isobaric mode by switching to polybaric mode and varying the minimum and maximum pressure of melting, while keeping the same average pressure (Table S2). This manual search, in the absence of independent constraints on the melt fraction, thus leads to the identification of several possible mantle sources for each target composition. The non-uniqueness and model uncertainty are discussed in more detail for the Fastball sample (see Supporting Information S1). We performed ~500,000 MAGMARS calculations by randomly varying the parameters around their average values. This automated search identified slightly larger compositional trends compared to the manual search. However, the mantle sources identified manually were found sufficient to discuss the mantle source origin and melting temperature. It is this data set (Table S2) that is described in the following sections.

### 4. Results

The compositions of the mantle sources that can produce melts identical to the target compositions (Table 1 and Figure 1) are shown in Figure 2 and reported in Table S2. Each target composition can be matched by melting a series of mantle sources characterized by various concentrations of incompatible elements (Al, Ti, Ca, Na, K, and P), both isobarically and polybarically. Despite the non-uniqueness of solutions, first-order chemical differences between the sources of the different basaltic compositions can be identified. For example, the possible sources of shergottites are all notably poorer in  $Al_2O_3$  and  $Na_2O$  than the sources of Gusev basalts (Figure 2a). Among the latter, the sources of Columbia Hills basalts are characterized by high  $Na_2O$ ,  $K_2O$ , and  $P_2O_5$  concentrations (Figures 2b–2d) compared to the source of Adirondack basalts. The source of the NWA 2737 chassignite shows the highest  $K_2O/Na_2O$  ratio. Finally, source 1 of Clast VI (NWA 7533) is nearly identical to the DW85 primitive mantle.

The melt fractions required to produce the target compositions are comprised between 5 and 30 wt.%. The associated mantle potential temperatures ( $T_p$ ) are between 1320 and 1520°C (Figure 3a and Table S2). The average pressure of melting is relatively low for all samples (1.1–2.0 GPa), and is largely constrained by the  $SiO_2$  and MgO concentrations of the target primary melts (Figure 1a). If a Mg# of 81 (K22) is assumed instead of 77 for NWA



**Figure 1.** Comparison between the composition of Martian primitive basalts (left extremity of colored lines), their recalculated primary melts (symbols) and the melts produced by melting of the primitive mantle of Dreibus and Wänke (1985) at 1.5 (gray line) and 3.0 GPa (black line), as calculated by MAGMARS (Collinet et al., 2021). The high  $\text{SiO}_2$  content of primary melts (a) is consistent with shallow melting conditions ( $\ll 3.0$  GPa). But compared to shallow DW85 melts (1.5 GPa), many primary basalts have either lower or higher CaO contents (b) and higher  $\text{Na}_2\text{O}$  and  $\text{K}_2\text{O}$  contents (c–d), and must therefore derive from mantle sources of contrasting compositions.

7034/7475/7533 and Gusev basalts, then the primary melts would have fractionated more olivine and the mantle  $T_p$  (1390–1570°C) and average pressure of melting (1.9–3.0 GPa) would both be higher (Figures 3a and 3b).

## 5. Discussion

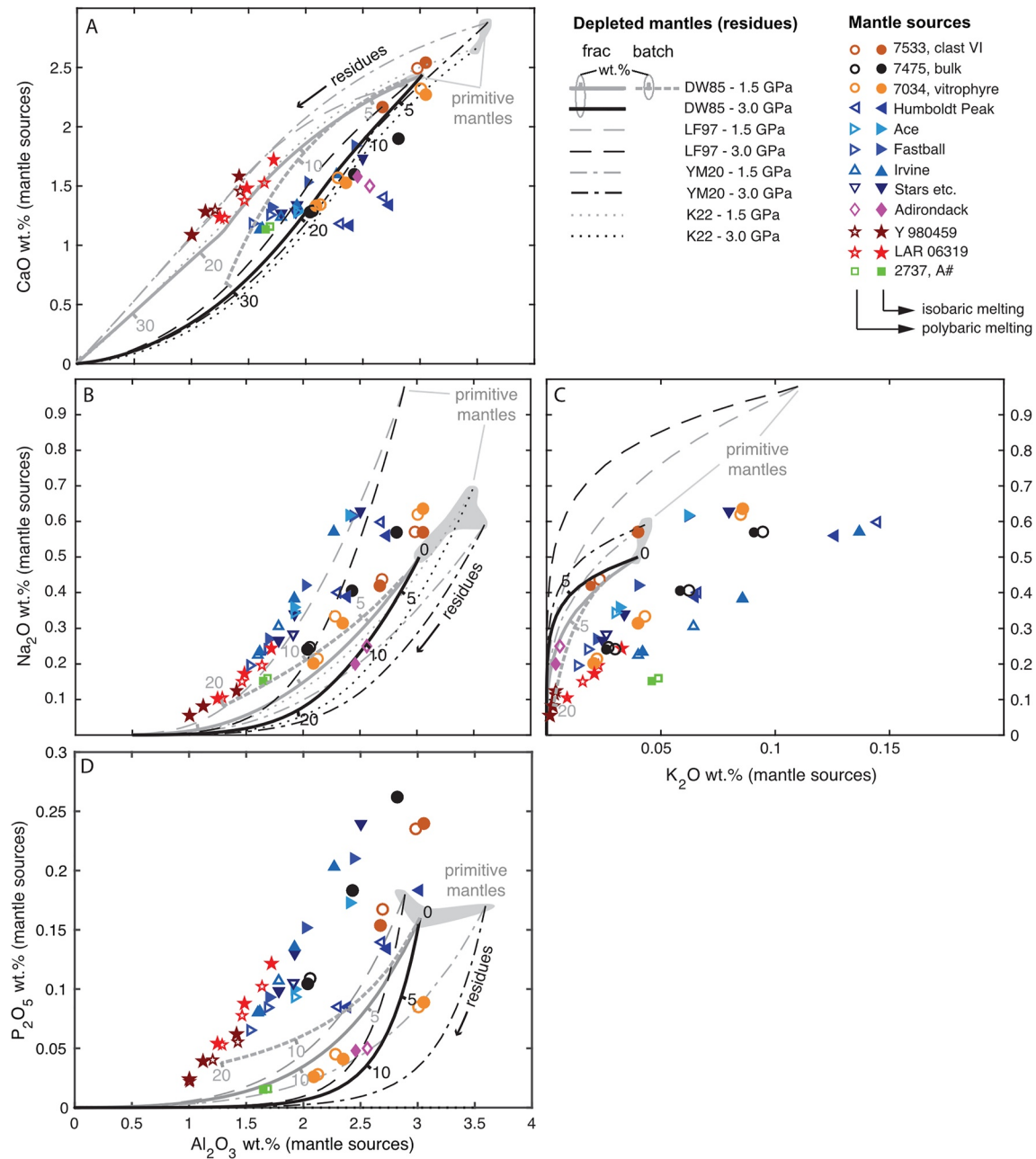
### 5.1. Thermal State of the Martian Mantle

Compared to the  $T_p$  estimates of Filiberto (2017), and using the same starting assumptions (mantle of Mg# 77 and batch melting), we find that Gusev crater basalts are derived from slightly cooler mantle sources on average, with  $T_p$  of  $\sim 1400^\circ\text{C}$  (vs.  $\sim 1450^\circ\text{C}$ ), but that the ranges of possible  $T_p$  largely overlap (1360–1460 vs. 1390–1550°C, respectively; see also Table S4). Allowing for a higher Mg# of the mantle sources (77–81), we find that Gusev basalts and all (pre-)Noachian to Hesperian samples point to a  $T_p$  of 1340–1520°C (Figure 3a).

We calculate a  $T_p$  of 1420–1430°C for the primary melt composition reconstructed from NWA 2737 melt inclusions (He et al., 2013), assumed to be parental to the middle-Amazonian nakhlites and chassignites (1.34 Ga; Udry & Day, 2018). However, the mantle source could have been metasomatized (Day et al., 2018, also see Section 5.2) and could have contained up to 250 ppm of water (McCubbin et al., 2016), which would translate into a lower  $T_p$  of 1380°C (Katz et al., 2003).

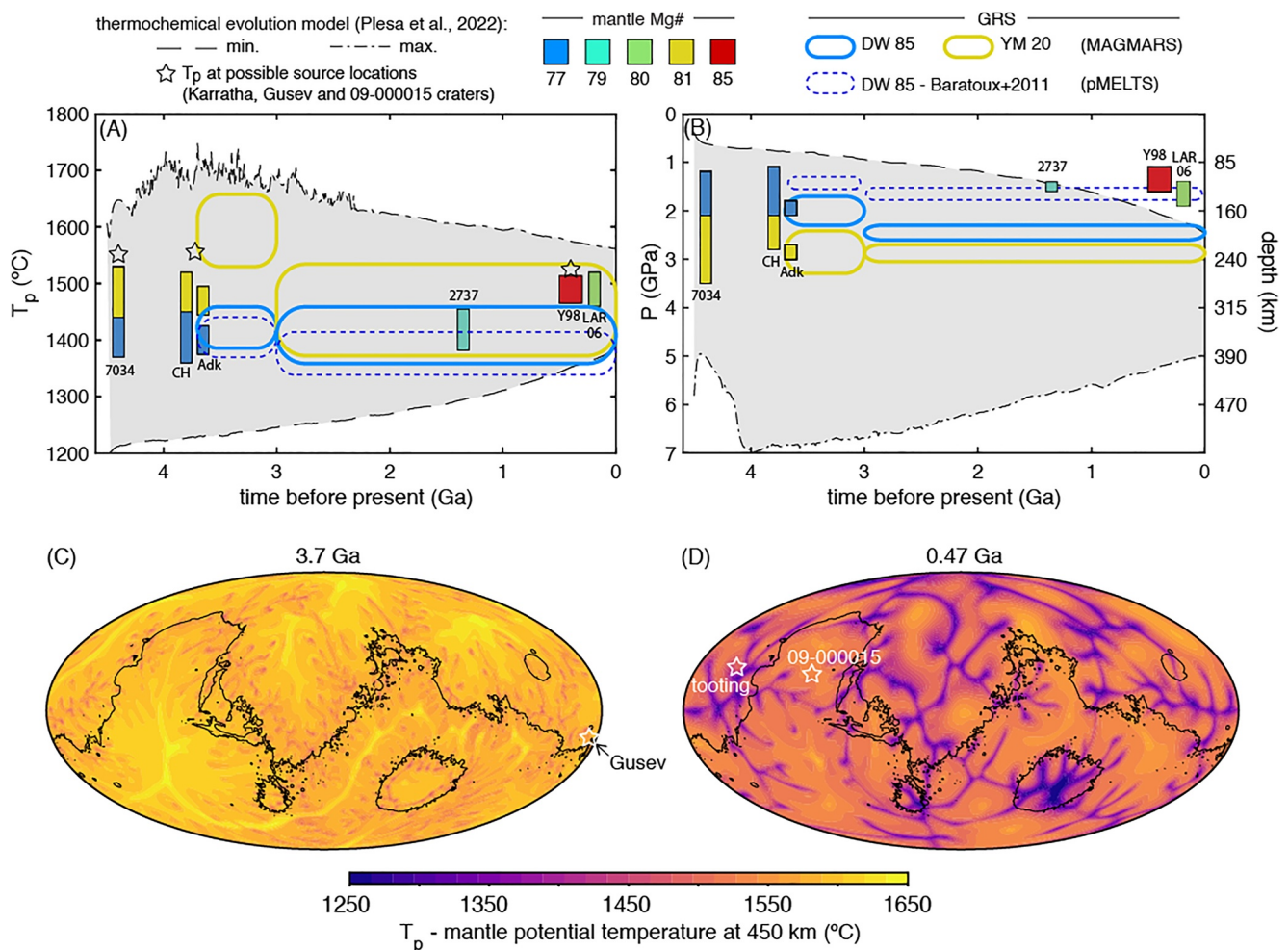
The more recent olivine-phyric shergottites (160–500 Ma; Moser et al., 2013; Wu et al., 2021; McFarlane & Spray, 2022) are picritic basalts that have been linked to plumes with a  $T_p$  of at least 1480–1550°C (e.g., Filiberto & Dasgupta, 2015; Musselwhite et al., 2006). The  $T_p$  of MAGMARS simulations (1470–1520°C) are within error of these previous constraints if batch melting is assumed, and slightly lower in the polybaric case (1440–1450°C).





**Figure 2.** Incompatible element concentrations of the mantle sources of primary basalts (symbols) compared to residual model Martian mantles (lines). Each line represents the trajectory of residues produced by progressive melting of a primitive mantle composition (apex) at 1.5 (gray) and 3.0 GPa (black). For the DW85 model (solid lines), tick marks indicate the composition of residues after specific degrees of melting (in wt.%). All mantle source compositions are relatively poor in CaO and  $\text{Al}_2\text{O}_3$  compared to the primitive mantle and are characterized by variable  $\text{CaO}/\text{Al}_2\text{O}_3$  (a). The source of NWA 7533 clast VI is similar to a primitive mantle (DW85) and the source of the Adirondack basalts resemble a residual mantle following prior partial melting in all compositional spaces (a–d). All other sources are too rich in alkali elements— $\text{Na}_2\text{O}$  (b) and especially  $\text{K}_2\text{O}$  (c)—and other incompatible elements, such as  $\text{P}_2\text{O}_5$  (d), to derive from model Martian mantles by partial melting alone and other processes must be considered (see Section 5.2).

Finally, using MAGMARS, we re-calculate the  $T_p$  and pressures of melting of the bulk volcanic provinces of Baratoux et al. (2011), as constrained by the Gamma Ray Spectrometer (GRS) on board NASA's Mars Odyssey spacecraft. Baratoux et al. (2011) used pMELTS in their analysis, which has since been shown to overestimate FeO and underestimate  $\text{SiO}_2$  concentrations by up to 8 wt.% (Collinet et al., 2021), significantly more than anticipated by El Maarry et al. (2009). For Hesperian provinces, while the ranges of  $T_p$  are similar (1390–1460 vs. 1370–1420°C previously), MAGMARS predicts a slightly higher pressure of melting (1.6–2.3 vs. 1.3–1.6 GPa).



**Figure 3.** Temporal evolution of  $T_p$  (a) and the average pressure and depth of melting or conditions of mantle–melt re-equilibration (b). The rectangles represent the sources of the basaltic compositions listed in Table 1. The rounded fields are the sources of the Gamma Ray Spectrometer volcanic provinces of Baratoux et al. (2011), re-calculated with MAGMARS. The black lines represent the evolution of the potential temperatures and pressures of the part of the mantle that is affected by partial melting in the thick-crust geodynamical model of Plesa et al. (2022). The minimum pressure of melting (dashed line in B) can be interpreted as the thinnest thermal lithosphere observed anywhere on the planet. Panels (c) and (d) represent regional variations in  $T_p$  for this geodynamical model at the time of Gusev basalt (c) and depleted shergottites (d) crystallization. At their possible source locations (white stars, see text for references), the  $T_p$  are nearly identical: 1562 vs. 1525°C (see also Figure S3 in Supporting Information S1).

However, we find that only Ascraeus and Elysium Mons (out of the 6 Amazonian volcanic provinces) can be matched with a DW85 mantle composition using MAGMARS (Table S3). The composition of the other 4 provinces can either not be reproduced at all (Arsia and Pavonis Mons) or only with an extremely small melt fraction of <2 wt.% (Olympus Mons and Alba Patera). With a YM20 composition (Mg# of 79, 81 after 15 wt.% of melting), a higher  $T_p$  of 1520–1660°C and higher pressures of melting (2.3–3.5 GPa) are necessary to match the Hesperian volcanic provinces (Figure S4 in Supporting Information S1). A higher Mg# mantle also allows to reproduce the composition of a greater number of Amazonian volcanic provinces (5, all but Arisia Mons) with  $T_p$  of 1380–1460°C and pressures of 2.8–3.1 GPa.

The lack of temperature and pressure trends over time displayed by this set of constraints renders it impossible to calculate rates of secular cooling or lithosphere thickening (Figures 3a and 3b). This could be due to the limited number of primitive basalts available that might not be representative of the average mantle. To test this possibility, we compare the mantle temperature estimates derived from MAGMARS to the results of a global thermochemical evolution model incorporating the most recent interior structure constraints from InSight (Plesa et al., 2022). This model predicts that the average mantle temperature—and maximum temperature at which basaltic melts can be produced (Figure 3a)—should first increase due to the decay of radioactive elements and

peak at 4.0–3.5 Ga before slowly decreasing. By inspection of the spatial variability of the mantle temperature, we observe that melt is produced in regions of the mantle with highly variable  $T_p$  at any given time (Figures 3a, 3c, and 3d). The Gusev basalts could thus simply derive from a relatively cold region of the mantle, while NWA 7034 and the depleted shergottites could derive from regions of the mantle that were close to the maximum  $T_p$ , at the time at which they crystallized (assuming that the Karratha and 09-000015 craters are the respective ejection sites; Lagain et al., 2021; Lagain et al., 2022). On the other hand, our re-interpretation of the  $T_p$  of Hesperian volcanic provinces (1520–1660°C, for a mantle Mg# of 79; Baratoux et al., 2011) is in line with the prediction that the mantle of Mars was, on average, substantially warmer during the Hesperian than during the Amazonian.

Perhaps the main discrepancy between the thermochemical evolution model and the MAGMARS constraints is the shallow depth of melting that we estimate for the source of shergottites, which is predicted to be well within the lithospheric mantle (Figure 3b). Filiberto (2017) noted that if a larger amount of olivine fractionation had taken place, the primary melts of shergottites could have been in equilibrium with the convecting mantle at 3–5 GPa. While this pressure of melting is more consistent with the thick lithosphere of the late Amazonian (Figure S3 in Supporting Information S1), such melt compositions would require a high  $T_p$  of  $1710 \pm 73^\circ\text{C}$ , which exceeds significantly the maximum  $T_p$  achievable by thermal evolution models at that time (Figure 3d). Therefore, we consider it more likely that the  $T_p$  of the sources was low (1470–1520°C) and that the pressure derived from MAGMARS simulations ( $1.6 \pm 0.5$  GPa) does not represent the average pressure of melting but simply the final pressure of equilibration with the mantle. If shergottites formed in the Tharsis region (e.g., Lagain et al., 2021), deeply-sourced primary melts could have re-equilibrated with a warm lithospheric mantle, locally heated by magmas, at the base of the crust (110–130 km; Wieczorek et al., 2022). Alternatively, the presence of water could lower the required  $T_p$  and help account for the relatively high  $\text{SiO}_2$  concentrations of shergottites (Balta & McSween, 2013b). However, the small initial water concentration that has been suggested for their source (14–73 ppm; McCubbin et al., 2016) does not favor the formation of shergottites from hydrous melts.

## 5.2. Origin of the Mantle Sources and Their Variable Concentrations of Incompatible Elements

The mantle source of Clast VI (NWA 7533) could be nearly identical to the primitive mantle (Figure 2; source 1 in Table S2), as previously suggested based on rare-earth element modeling (Humayun et al., 2013). All other mantle sources are depleted in CaO and  $\text{Al}_2\text{O}_3$  relative to the various primitive mantle compositions proposed in the literature (DW85, K22, LF97, and YM20). One possibility is that these mantle sources represent melting residues of the primitive mantle from which 10–20 wt.% melt were removed prior to the formation of the target basaltic compositions used in this study (Figure 2a). However, the concentrations of alkalis and other incompatible elements (e.g.,  $\text{TiO}_2$ ,  $\text{P}_2\text{O}_5$ ) are, in most cases, too high at a given  $\text{Al}_2\text{O}_3$  concentration, regardless of the style (batch vs. fractional) and pressure of melting (Figures 2b–2d). Only the Adirondack basalts are consistent in detail with the simple re-melting of a mantle residue, following  $\pm 10$  wt.% prior melting of a primitive mantle (see also Collinet et al., 2021). Other processes must be invoked to explain the chemical variability of the remaining mantle sources.

The Columbia Hills basalts are often assumed to be related to the Adirondack basalts, as both groups were analyzed by *Spirit* at Gusev crater. Compared to the Adirondack basalts, they are rich in alkali elements as well as other incompatible elements ( $\text{TiO}_2$ ,  $\text{P}_2\text{O}_5$ ) and poor in CaO and  $\text{Al}_2\text{O}_3$  (Figure 1). McSween, Ruff, et al. (2006) suggested that the Columbia Hills basalts could have derived from melts similar to the Adirondack basalts by fractional crystallization. The higher incompatible element concentrations (e.g., K, P, Ti) of the Columbia Hills basalts have also been suggested to result from the contamination of Adirondack-like primitive melts by a crustal component (Schmidt & McCoy, 2010). However, crustal assimilation and fractional crystallization of basaltic melts should lower markedly the MgO concentrations (and Mg#; Ostwald et al., 2022). As the Mg# of the Columbia Hills and Adirondack basalts are similar, most workers now regard them as two sets of near-primary melts (Collinet et al., 2015; Filiberto & Dasgupta, 2011; Schmidt & McCoy, 2010). Schmidt and McCoy (2010) proposed that the high  $\text{K}_2\text{O}$  content of the Columbia Hills basalts could be accounted for by melting a fertile mantle source with a higher  $\text{K}_2\text{O}$  content compared to the Dreibus and Wänke (1985) composition. According to their model, the Adirondack basalts would be slightly younger and produced by re-melting the same region of the mantle. However, the similarly low CaO and  $\text{Al}_2\text{O}_3$  concentrations of their sources (Figure 2a) suggest that both the Adirondack and Columbia Hills basalts were derived from depleted mantles, affected by 10–20 wt.% prior melting at  $\sim 3.0$  GPa. Metasomatism has been invoked to reconcile the high water and incompatible element

concentrations of nakhlites-chassignites with their Sr-Nd isotopic compositions indicative of ancient depleted sources (Day et al., 2018; Goodrich et al., 2013; McCubbin et al., 2013) and could also help explain the high K<sub>2</sub>O concentrations analyzed in numerous rocks from Gale crater (e.g., Schmidt et al., 2014). Similarly, we posit that the relative enrichment of incompatible elements in the Columbia Hills basalts (alkali elements as well as elements like P and Ti that are less mobile in fluids) could be explained by the secondary addition of low-degree melts to a Adirondack-like mantle source. The highest possible K<sub>2</sub>O concentrations that we calculate for the Columbia Hills mantle sources are in the range 0.13–0.15 wt.%. This is much smaller than the percent level K<sub>2</sub>O concentrations of highly metasomatized and phlogopite-bearing terrestrial peridotites (e.g., Condamine & Médard, 2014) but similar to other intraplate peridotites containing no hydrous phases (e.g., Smith et al., 1993). The source of the Columbia Hills basalts was likely affected by low degrees of cryptic metasomatism and was thus not significantly hydrated.

The isotopic systematics of Martian meteorites suggest the existence of a magma ocean that crystallized early in Mars' history (e.g., Bouvier et al., 2018; Debaille et al., 2008; Elkins-Tanton et al., 2005; Kruijer et al., 2017). Some of the resulting heterogeneity was never erased by convection and ancient mantle sources were affected by partial melting and formed the shergottites as recently as 170 million years ago (McFarlane & Spray, 2022; Moser et al., 2013; Wu et al., 2021). The major and incompatible element concentrations of the sources of shergottites must in part reflect the processes of magma ocean crystallization. For example, the superchondritic CaO/Al<sub>2</sub>O<sub>3</sub> ratio of shergottites has been suggested to result from the fractionation of majorite in the deep mantle (Borg & Draper, 2003). Here, we find that the sources of shergottites had mildly superchondritic CaO/Al<sub>2</sub>O<sub>3</sub> ratios that could have appeared at low pressure, following 15–20 wt.% melting of the primitive mantle (Figure 2a). A 20 wt.% depletion from a primitive mantle is also sufficient to decrease the incompatible element concentrations to levels identical to those of the source of depleted shergottites (Figures 2b–2d). In this case, however, the melting residue only reaches a Mg# of 77 (when starting from a DW85 mantle) to 81 (YM20), following 20 wt.% of melting. The much higher Mg# of the source of Y 980659 (85–86) remains easier to explain if it formed as a magma ocean cumulate (e.g., Borg & Draper, 2003; Elkins-Tanton et al., 2005). The enriched shergottites have higher concentrations of incompatible elements. Their composition in radiogenic isotopes indicates that the enriched signature is most likely derived from evolved residual melts that were trapped in mantle cumulates during the crystallization of an early MMO, rather than from crustal assimilation (e.g., Armytage et al., 2018; Borg & Draper, 2003; Brandon et al., 2012; Debaille et al., 2008; Ferdous et al., 2017; Symes et al., 2008). This could also explain the slightly higher concentrations of minor incompatible elements that we calculate for the source of enriched shergottites (Figure 2).

## 6. Conclusions

The mantle temperature of the sources that gave rise to known primitive basalts appears to have remained relatively stable through time ( $T_p$  of 1400–1500°C) but is likely due to a sampling bias. The higher mantle  $T_p$  (~1600°C) of the Hesperian volcanic provinces (Baratoux et al., 2011), recalculated with MAGMARS and assuming a mantle of Mg# 79 or higher (Khan et al., 2022; Yoshizaki & McDonough, 2020), hint at a significant secular cooling (>100°C), as expected from thermochemical evolution models (Plesa et al., 2022). The shergottite melts were likely produced at pressures greater than 3 GPa but re-equilibrated with the lithospheric mantle at 1–2 GPa, for example, at the base of the thick Tharsis crust.

With the exception of the source of NWA 7034 and paired rocks, the mantle sources of known Martian basalts were poorer in Al<sub>2</sub>O<sub>3</sub> and CaO compared to primitive mantle compositions (e.g., Dreibus & Wänke, 1985; Yoshizaki & McDonough, 2020). The compositions of the sources of Gusev crater basalts that we calculate do not explicitly require a magma ocean stage and could represent simple depleted mantle reservoirs affected by 10–20 wt.% prior melting (Adirondack basalts) or depleted mantle reservoirs re-fertilized by fluids and low-degree silicate melts (Columbia Hills basalts). On the other hand, the major element composition of the source of depleted shergottites cannot be easily explained by partial melting alone and suggest, along with their Sr-Nd-Hf isotope systematics, that they formed as mantle cumulates during the crystallization of the MMO. The concentration of major and minor incompatible elements in the source of enriched shergottites is consistent with the trapping a more evolved residual melt.

## Data Availability Statement

The data used for the discussion and figures is summarized in Tables S1–S4 and available in full at <https://doi.org/10.5281/zenodo.7949084> (Collinet et al., 2023).



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