A simple chondritic model of Mars

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Abstract

SNC meteorites (Shergottites, Nakhlites and Chassigny) define a fractionation line in a δ17O/δ18O diagram as expected for rocks differentiated from a formerly homogenized parent body, which is intermediate between H ordinary chondrites and EH enstatite chondrites. The planet Mars is located between the Earth and the asteroid belts, the potential source of ordinary chondrites. Since oxygen is a major component of the terrestrial planets, our model assumes that Mars composition is a mixture of two chondritic sources whose proportions are calculated by mass balance based on oxygen isotopes only. Two possible model compositions can be derived: if the average isotopic composition of SNC is relaxed along its fractionation line (Model 1) and if the end members are average H and EH chondrites, one obtains a 30:70 H:EH mixture. If the average isotopic composition of SNC is a robust feature, then an extreme composition of H chondrites must be selected which yields the proportion, 55:45 for the H:EH components. This composition carries the same oxygen isotopic composition as the iron inclusions in the IIE of the conjectured end member of the ordinary chondrite group. The proportions obtained this way enable to calculate two model compositions for all the refractory elements and oxygen. Model 1 can be discarded as it does not permit to fit reasonably the physical properties of the planet. Mass and composition of the core (Model 2) is easily derived (23% of Mars mass, containing 16% Sr); the remainder forming the bulk mantle composition. Comparison with recent estimates based on the composition of SNC meteorites reveals only minor differences, essentially for Si, Mg and Fe; this is because of our choice of non-CI chondritic composition, unlike previous models. Discussion of the assumptions made in previous models confirms that the new composition is in agreement with the SNC compositions. The model also permits to calculate adequate physical properties of the planet like its zero pressure mantle density, density profile with depth and dimensionless moment of inertia. The superiority of the present model resides in the minimal number of necessary hypotheses and the possibility to test it, using physical and chemical data about the planet: as far as we know of, all constraints can be satisfied within the errors of the measurements or uncertainty of the model. The Fe abundance and the low Mg/Sr ratio imply that pyroxenes and garnet at depth will play a major role in the differentiation of the planet, a feature which differs markedly from the terrestrial mantle. © 1999 Elsevier Science B.V. All rights reserved.

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1. Introduction

In recent years, a number of works have been devoted to establish a bulk chemical composition of
Mars, even more since the Martian origin of the SNC meteorites (Shergottites, Nakhlites and Chassigny) has been recognized. These attempts for reconstructing compositional and/or physical models of the planet were constrained by a limited number of information: (1) the physical data: mass, size and moment of inertia (Anderson, 1972; Moquet et al., 1996; Bertka and Fei, 1997, 1998a,b); (2) the conjecture that the SNC meteorites are fractionated rock ejecta from the planet (McSween, 1985); and (3) the reference to chondritic meteorites, assuming that most if not all terrestrial planets are derived in some way from similar material (Dreibus and Wänke, 1985; Laul et al., 1986; Longhi et al., 1992).

Additional inputs are necessary to solve the problem: (4) the presently most elaborated model (Dreibus and Wänke, 1985; Longhi et al., 1992) assumes that the major refractory elements occur with CI carbonaceous chondritic abundance ratios; and (5) volatile elements are derived from elemental correlations among SNCs, assuming simple behavior during differentiation as is observed for terrestrial or lunar samples (Dreibus and Wänke, 1985).

This kind of approach is certainly valuable, but we see a number of difficulties. (i) The model involves as many hypotheses as unknown. All available information (chemical or physical) for the planet are necessary inputs to set up the model with the consequence that no independent observations can be used to check for its validity. (ii) The abundance of sulfur, a critical number because of its implication on the size and density of Mars core is fixed at $0.35 \times \text{CI}$ value, based on the supposedly comparable volatility of moderately volatile elements (e.g., Fe, Ga, Na, P, K, F and Rb). This could be considered as an ad hoc hypothesis, since sulfur volatility is strongly dependent on its oxidation state (see next point). (iii) The abundance of the most abundant element on Mars, oxygen, is not specified by the previous models. This point we consider of utmost importance, since the core size depends chiefly on the oxidation state of iron. Different classes of chondrites contain variable amounts of oxygen and this is actually one of the criteria used for their classification. The same is true for the terrestrial planets according to the size of their metallic (±sulfide) core. This observation contrasts with the fully oxidized character of CI chondrites. In previous models, the size of the Martian core was constrained by the mass and moment of inertia of the planet. Again, if an independent estimate of O abundance was available, then the physical properties could be used to test the model validity.

We therefore propose a new approach, considering that a major constraint was disregarded in the previous models. Oxygen is the most abundant atom in the planet, but surprisingly, neither its abundance (equivalent to the ratio of reduced to oxidized iron) nor its isotopic composition, have been considered in previous models. Still, it can be read in textbooks that these two parameters bring important constraints for the classification of meteorites in relation with the early structure and temperature of the solar nebula. The isotopic composition of SNC meteorites (Clayton and Mayeda, 1983, 1996; Clayton et al., 1991; Romanek et al., 1998) is intermediate between that of enstatite chondrites and ordinary chondrites (Clayton et al., 1991) (Fig. 1). Mars orbits between the Earth (with oxygen isotopes similar to enstatite chondrites) and the asteroid belts, the potential source of ordinary chondrites (Wetherill, 1985). It is therefore appealing to consider Mars as a mixture of the two meteorite types without the need to make further hypotheses. Most previous models considered CI as the basic material to form planetary bodies like the Earth or Mars. As pointed out repeatedly (Smith, 1982; Javoy and Pineau, 1983; Javoy, 1995), their oxygen isotopes ($\delta^{18}O$ in the range 1.5–2.5%) cannot be reconciled with either Earth or Mars composition and their oxidation state raises insuperable difficulties in forming massive metallic cores. More recently, Lodders and Fegley (1997) have proposed a more complex model based on a mixture of CI, CV and H meteorites with the aim of reproducing the oxygen isotopic composition of SNC meteorites.

The aim of this work is as follows: considering the oxygen isotopic composition of the SNC, we calculate the appropriate mixture of chondritic meteorites necessary to reproduce their isotopic composition. This simple hypothesis does not require any knowledge of the moment of inertia and, except for oxygen isotopes, makes no reference to SNC chemical composition. The outputs of the model are the bulk chemical composition of the planet, including core size and composition and its physical properties (density profile, moment of inertia).
All the observations on Mars composition and physical properties will be used in a second step to check for the model validity, a possibility not offered by previous works. At this stage, it can be pointed out that the recent development of a terrestrial composition derived from EH chondrite material (Javoy, 1995, 1998) is a very similar approach. The former attempts of Ringwood (1981) and Wänke and Dreibus (1988) to construct a Martian composition from two component mixtures could also be viewed as a comparable approach; their models, however, were rather dictated by the failure of obtaining an appropriate composition derived from CI chondrites alone and simple processes, with a number of internal inconsistencies. Their two components were not identified with chondritic classes, but conveyed some characteristics of both EH and H chondrites; therefore, even though their pioneering works cannot be identified with the present model, some of their observations are implicitly contained in the present approach.

2. Mars as a mixture of chondritic meteorites: the case of oxygen isotopes

2.1. SNC isotopes relative to chondrites

The analyses of oxygen isotopes in meteorites have shown that the most important classes, defined on the basis of their chemical and mineralogical features can be reproduced according to their isotopes. In a $\delta^{17}\text{O}/\delta^{18}\text{O}$ diagram, conventional fractionation of isotopes may spread out an initially homogeneous reservoir along a straight line with a slope of 1/2 (more precisely 0.52). The other important characteristics in this diagram is that any mixing line is a straight line passing through the end members. As an example, all terrestrial samples fall along the Terrestrial Fractionation Line (TFL) passing through the origin (the reference standard is a terrestrial sample). Enstatite chondrites (EH and EL), aubrites and the Moon fall on the same line. Considering that this feature cannot be coincidental (Javoy,
1995, 1997), elaborated compositional models for the Earth based on an EH bulk composition.

Interestingly, this approach opens new perspectives about the upper and lower mantle differentiation, light element in the core and Mg/Si ratio of the mantle.

The oxygen isotopic compositions of the different meteorite classes are fairly well-known (Clayton and Mayeda, 1984; Clayton et al., 1991) as well as for the SNC meteorites. The latter fall, within error, along a fractionation line in a $^{17}\text{O}/^{18}\text{O}$ diagram (Fig. 1) which permits to exclude some achondrites from the SNC group and confirm the consanguinity of the others. A number of samples, whole rocks and separates, have been analysed which permit to define quite accurately the fractionation line with $\delta^{18}\text{O}$ ranging from 2.5 to 14.4$_{\text{o}}$ (Clayton and Mayeda, 1983, 1996; Romanek et al., 1998). An average $\delta^{18}\text{O}$ of 4.45$_{\text{o}}$ can be calculated from the whole rock compositions, but this value may vary slightly, depending on the weighting of the data points. More important is the position of the fractionation line in the diagram as it carries the parental signature of the planetary object. It clearly falls in-between the terrestrial fractionation line and the ordinary chondrite field. Ordinary chondrites (H, L and LL) display higher $\delta^{17}\text{O}$ values. They do not define a fractionation line, but their $\delta^{17}\text{O}/\delta^{18}\text{O}$ correlation can be described as a mixing line over, at least, two $\delta^{18}\text{O}$ units.

2.2. The chemical model

Therefore, we formulate the central hypothesis of the present model: Mars is a mixture of two components, one having an EH-like composition, the other an ordinary chondrite composition. Should ordinary chondrite composition fall on a fractionation line, then the mixing equation would have a unique solution without need to determine the average composition of either SNC or the end members. This follows from the simple geometrical property that parallels cut proportional segments on any secant line.

Unfortunately, ordinary chondrites form a mixing trend significantly distinct from a fractionation line. In such a case, the solution to the mixing problem is not unique: every member of the ordinary chondrite group would yield a particular solution. Our problem therefore consists in selecting the appropriate composition in the ordinary chondrite group.

First of all is to be noticed that the average $\delta^{17}\text{O}$ composition of both EH and SNC are less than 3$_{\text{o}}$. This implies that any likely end member from the ordinary chondrite group must belong to the H chondrites subgroup. Still, the choice among the H chondrites is not strongly constrained and we thus consider two possibilities.

2.2.1. Model 1: average H chondrites as one end member

The most obvious possibility consists in selecting the ordinary chondrite end member with an average H composition. The tie line which ensues cannot pass through both the average SNC and average EH values (Fig. 1). This is, however, an acceptable possibility if the Martian value differs from the average SNC. The average value obtained from the SNC carries a limited statistical significance, since only 12 objects have been analysed and the appropriate weighting of the objects in calculating the true average is not known. The intersection occurs at a $\delta^{18}\text{O}$ of 4.9$_{\text{o}}$ compared to the average SNC of 4.42$_{\text{o}}$ ± 0.3 (1$\sigma$). Obviously still, the likelihood of this model will increase with the closeness of the two values. As SNC data points fall between the EH chondrites and H chondrites, we may write mass balance equations expressing that model Mars is a mixture of EH and H which must be valid for all elements including oxygen and its isotopes. The first relationship expresses the bulk oxygen mass balance

\[ (\text{O}_{\text{Mars}}) = x(\text{O})_{\text{EH}} + (1-x)(\text{O})_{\text{H}}, \]  

where $x$ is the mass fraction of EH chondrites in the model mix and (O) is the mass fraction of oxygen in component i. A second equation expresses the mass balance for $^{18}\text{O}$

\[ (^{18}\text{O}/^{16}\text{O})_{\text{Mars}} = x(^{18}\text{O}/^{16}\text{O})_{\text{EH}}(^{16}\text{O})_{\text{EH}} \]
\[ + (1-x)^{18}\text{O}/^{16}\text{O}_{\text{H}} \times (^{16}\text{O})_{\text{H}}, \]

since $^{16}\text{O}$ represents 99.8% of all oxygen. This can be rewritten using the more conventional $\delta$ notation

\[ (\delta^{18}\text{O})_{\text{Mars}} = x(\delta^{18}\text{O})_{\text{EH}}(^{16}\text{O})_{\text{EH}} \]
\[ + (1-x)(\delta^{18}\text{O})_{\text{H}}(^{16}\text{O})_{\text{H}}. \]
The mass balance calculation is illustrated in Fig. 1. Since ordinary chondrites do not fall on a fractionation line, the solution of the mass balance equation will depend critically on the average H chondrite estimate.

First of all, an accurate oxygen content of the relevant chondrites is necessary in order to solve 1, 2 and 3. For ordinary H chondrites, the values given by Wasson and Kalleymen (1988) are adopted without further refinement. If the model Mars composition is located at the intersection between the EH:H line and the SNC fractionation line, we obtain an H:EH mix of 30:70 (Table 1). Notice that the choice of average EH composition as the second end member is not necessary. Therefore, there is no need to discuss the isotopic composition of the average EH value, but their oxygen content is necessary. For EH enstatite chondrites, all components listed by Wasson and Kalleymen add up to 95.2% only. We have no explanation for this difficulty, but this may be related to late alteration of these highly reduced meteorites or to poor consistency of the data sets used to calculate the average. In the absence of a better explanation, we normalized to 100% without further modification, i.e., without changing the proportion of oxygen to other elements.

2.2.2. Model 2: the tie line must pass through SNC and EH averages

In the first model, average H, SNC and EH are not aligned. We assumed that this can be accounted for by the uncertain estimates of the averages as above. Still, another point of view can prevail. It could be argued that calculating an average H chondrite composition is convenient, but meaningless. There is a continuous evolution from H to L to LL compositions and dividing the group in subclasses is rather artificial even though their compositions differ significantly. In other words, there is no constraint at all for preferring an average H composition or any other composition. Unlike ordinary chondrites, both SNC and EH fall on a fractionation line which can be understood as resulting from the differentiation from an initially homogeneous chemical reservoir. It can therefore be postulated that for the latter, calculating an average composition is justified and probably very close to the exact (unknown) average value. Accepting their average compositions (within uncertainty) and selecting the appropriate H chondrite composition to fall on the mixing line so defined permits to solve the problem in a different way. Here we consider that the H chondrite population is heterogeneous and that the appropriate end member in the ordinary chondrite group has to be specified. The present difficulty consists in having the best estimate of average EH (isotopic composition and oxygen abundance) and SNC (isotopes only). We have no strong argument in selecting the best EH end member (EL can be discarded at once because of their too low iron content). The average EH value at 5.4‰ intermediate between the aubrite and Earth–Moon estimate within about ±0.1‰ suggests a single unit above the SNC average. This gives an estimate of the uncertainty inherent to the model.

The average SNC composition has been calculated from the whole rock data of Clayton and Mayeda (1983, 1984, 1996). According to the work of Romanek et al. (1998), one result for Lafayette has been discarded as it revealed low-temperature iddingsitization. Only 12 SNC samples have been analysed, some of which several times (19 analyses). The average giving the same weight to all samples is 4.42 ± 0.3 (1σ). When calculated in different ways, it always remains within the range 4.37 to 4.47. In this second model, we obtain an H:EH mix of 55:45. The H composition obtained in this model may seem extreme (δ18O = 3‰) and therefore unlikely. There are two arguments against this view.

(1) If the mixture exhibited by the ordinary chondrite population results from some isotopic zonation in the solar nebula, then an extreme end member can be viewed as the necessary choice: ordinary chondrites probably originate from the asteroid belt, between Mars and Jupiter, while enstatite chondrites (isotopically similar to the Earth and Moon) originate from a more central region in the nebula. One
could still claim that the extreme composition obtained is unlikely because very few such chondrites have been analysed which are rather outliers than representative samples: such an extreme component may never have existed at all. This is probably not justified either.

(2) Clayton and Mayeda (1996) measured oxygen isotopes in the inclusions of iron meteorites. Fractionation lines are observed for some of the iron classes as for stony meteorites, in particular, IIE and IVa iron meteorites yield two extreme fractionation lines corresponding to two end members of the ordinary chondrite population (Clayton and Mayeda, 1996). The IIE group appears, from an isotopic point of view and within uncertainty, identical to our H end member. The important point is that such a composition cannot be refuted on the basis of its unrepresentativity. We believe, however, that the similarity between IIE isotopic composition and our proposed end member, cannot be coincidental and hence is another argument in favor of this model.

It is difficult to evaluate precisely the uncertainties in mixing proportions for the following reasons. A more precise determination of the end-members composition is prevented by the scatter in the isotopic compositions of the meteorites, which is related to analytical precision and presumed poor sampling of meteorites which could result in a statistical bias. Therefore, the model compositions are precise to within a few percent.

3. Composition of the planet

3.1. Major elements and core/mantle differentiation

Knowing the proportions of the end-members meteorites, the bulk model compositions of Mars can be calculated easily from the chondritic compositions as given by Wasson and Kalleymen (1988), using a simple mass balance equation. The major element results are given in Table 2 and will be discussed below. In particular, is there any way to choose between models 1 and 2? Another crucial point is the differentiation of the planet. We can calculate the mass fraction and composition of Mars’ core and mantle which are fundamental data for the understanding of the planet (Table 3).

<table>
<thead>
<tr>
<th>Element</th>
<th>EH</th>
<th>H</th>
<th>Bulk EH45:55</th>
<th>Bulk EH70:30</th>
</tr>
</thead>
<tbody>
<tr>
<td>Mg</td>
<td>11.08</td>
<td>13.85</td>
<td>12.6</td>
<td>11.9</td>
</tr>
<tr>
<td>Si</td>
<td>17.45</td>
<td>16.60</td>
<td>17.0</td>
<td>17.2</td>
</tr>
<tr>
<td>Al</td>
<td>0.85</td>
<td>1.11</td>
<td>1.0</td>
<td>0.9</td>
</tr>
<tr>
<td>Ti</td>
<td>0.05</td>
<td>0.06</td>
<td>0.06</td>
<td>0.05</td>
</tr>
<tr>
<td>Ca</td>
<td>0.89</td>
<td>1.23</td>
<td>1.1</td>
<td>1.0</td>
</tr>
<tr>
<td>Mn</td>
<td>0.23</td>
<td>0.23</td>
<td>0.23</td>
<td>0.23</td>
</tr>
<tr>
<td>Na</td>
<td>0.71</td>
<td>0.63</td>
<td>0.67</td>
<td>0.70</td>
</tr>
<tr>
<td>Cr</td>
<td>0.33</td>
<td>0.36</td>
<td>0.35</td>
<td>0.34</td>
</tr>
<tr>
<td>Fe</td>
<td>30.31</td>
<td>27.01</td>
<td>28.5</td>
<td>29.3</td>
</tr>
<tr>
<td>S</td>
<td>6.03</td>
<td>1.96</td>
<td>3.8</td>
<td>4.8</td>
</tr>
<tr>
<td>Ni</td>
<td>1.83</td>
<td>1.57</td>
<td>1.7</td>
<td>1.8</td>
</tr>
<tr>
<td>O</td>
<td>29.26</td>
<td>35.07</td>
<td>32.5</td>
<td>30.1</td>
</tr>
</tbody>
</table>

Data are in wt.%, normalized to 100%.

For all refractory elements, i.e., those having a 50% condensation temperature in excess of 1250 K (Grossman and Larimer, 1974), a model Mars composition can be inferred from simple mass balance calculations. As 44 out of 86 elements are refractory, this provides us with a significant data base without further consideration. Among the 14 most abundant elements, only two—Na and S—are moderately volatile and may cause problems. As Na is a lithophile element, its abundance will be derived from the composition of the SNC meteorites following Wanke and Dreibus (1988). Its low abundance, however, will not affect the properties of the planet and this point can be disregarded to begin with. The same is not true for S as it is far more abundant and expected to partition into the core as liquid sulfide. Its abundance cannot therefore be directly derived from the SNC which are differentiated objects. The effect of S abundance is not negligible as it will strongly influence the core density and its melting point (Usselman, 1975; Fei et al., 1997). Assuming no loss of moderately volatile elements from the EH and H chondritic precursors during the final stages of accretion, the S abundance corresponds to an equivalent of 40–50% CI abundance (Si normalized). When compared to the hypothesis of depletion-like elements of similar volatility (Longhi et al., 1992), suggesting that Mars’ sulfur content is about 0.35% CI abundance, one may conclude that S volatilization occurs mainly in the chondritic precursors them-
Table 3: Model compositions of Martian mantle and core (in wt.%, normalized to 100% for EH–H models)

<table>
<thead>
<tr>
<th>Model</th>
<th>EH45:H55</th>
<th>EH70:H30</th>
<th>Anderson, Wanke and Dreibus, 1972</th>
</tr>
</thead>
<tbody>
<tr>
<td>Mantle</td>
<td>0.76</td>
<td>0.72</td>
<td>0.88</td>
</tr>
<tr>
<td>MgO</td>
<td>27.3</td>
<td>27.3</td>
<td>27.4</td>
</tr>
<tr>
<td>SiO2</td>
<td>47.5</td>
<td>51.0</td>
<td>40.0</td>
</tr>
<tr>
<td>Al2O3</td>
<td>2.5</td>
<td>2.5</td>
<td>3.1</td>
</tr>
<tr>
<td>TiO2</td>
<td>0.1</td>
<td>0.1</td>
<td>0.1</td>
</tr>
<tr>
<td>CaO</td>
<td>2.0</td>
<td>2.0</td>
<td>2.5</td>
</tr>
<tr>
<td>MnO</td>
<td>0.4</td>
<td>0.4</td>
<td>0.2</td>
</tr>
<tr>
<td>Na2O</td>
<td>1.2</td>
<td>1.3</td>
<td>0.8</td>
</tr>
<tr>
<td>Cr2O3</td>
<td>0.7</td>
<td>0.6</td>
<td>0.6</td>
</tr>
<tr>
<td>FeO</td>
<td>17.7</td>
<td>11.4</td>
<td>24.3</td>
</tr>
<tr>
<td>Mg/Si (mole)</td>
<td>1.17</td>
<td>0.98</td>
<td>1.53</td>
</tr>
<tr>
<td>Mg#</td>
<td>0.72</td>
<td>0.80</td>
<td>0.67</td>
</tr>
<tr>
<td>Core</td>
<td>0.23</td>
<td>0.28</td>
<td>0.12</td>
</tr>
<tr>
<td>Fe</td>
<td>76.6</td>
<td>76.3</td>
<td>72.0</td>
</tr>
<tr>
<td>S</td>
<td>16.2</td>
<td>17.4</td>
<td>18.6</td>
</tr>
<tr>
<td>Ni</td>
<td>7.2</td>
<td>6.3</td>
<td>9.3</td>
</tr>
</tbody>
</table>

Anderson’s and Wanke and Dreibus’ models are displayed for comparison. In the EH–H models, the Na content is probably overestimated due to its volatility.

3.2. Minor elements in the mantle

Minor elements can be divided into three categories: (1) refractory lithophiles, (2) volatile lithophiles and (3) siderophiles (volatile or not). Determining the abundance of the 28 refractory lithophile elements is straightforward. It directly follows from the composition of the chondritic precursors; their abundance is listed in Table 2. Without further assumption, we thus obtain an estimate for 35 elements. For all other elements, the procedures proposed by Wanke and Dreibus (1988) can be used as they are probably the best, but as they strongly rely on the SNC composition, discussing their abundance falls beyond the scope of this paper.

4. Physical model

From the previously described chemical compositions of Mars and a physical model, it is possible to calculate the core size, the pressure and the density profiles of the planet. The physical model assumes a global hydrostatic state equilibrium

$$P(r) = \int_r^R \rho(x) g(x) dx$$ (4)
and includes two equations of state for pressure (one for the mantle, one for the core) of the following form

\[ P(r) = \frac{3K_0}{2} \left[ \left( \frac{\rho}{\rho_0} \right)^2 - \left( \frac{\rho}{\rho_0} \right)^5 \right] \times \left( 1 + \frac{3}{4} \left( K'_0 - 4 \right) \left( \frac{\rho}{\rho_0} \right)^2 - 1 \right) + \alpha K(T - 300) \]  

(5)

(the first term corresponds to a third-order Birch–Murnaghan isothermal room-temperature equation of state; the second term corresponds to thermal pressure, where \( K_0 \) is the zero-pressure bulk incompressibility, \( K'_0 \) its pressure derivative, \( \alpha \) is the thermal expansion coefficient and \( \rho_0 \) is the zero-pressure volumic mass).

We use an iterative procedure. Starting from an initial approximate density profile, a hydrostatic value of the pressure profile is calculated. The equations of state for the mantle and for the core (5) are then inverted to find a new density profile and both calculations are repeated until convergence is achieved. As two sets of parameters are used to solve Eq. (5) in the mantle and in the core, the core radius must be estimated. This value is used to solve Eq. 5 in the mantle and in the core, where \( M \) is the mass of Mars, 0.038 is its bulk sulfur content (EH45:H55 case), \( M_{c99} \) is the mass of the core minus its most external shell (for needs of calculus, we divided both mantle and core in 100 spherical shells) and \( r_{c99} \) is the radius of the 99th shell of the core. \( R_c \) is then calculated at each iterative step from the values of \( M_{c99} \), \( r_{c99} \) and \( \rho_{c100} \) taken at the previous step.

The core mass derived from our chemical model (Table 4), determines uniquely the zero-pressure mantle density from

\[ \frac{M_c}{M} = 1 - \frac{1}{M} \int_{R_c}^{R} 4\pi r^2 \rho_{mantle} \, dr, \]  

(7)

Table 4

<table>
<thead>
<tr>
<th>Physical properties of Martian models</th>
</tr>
</thead>
<tbody>
<tr>
<td>( R_c )</td>
</tr>
<tr>
<td>( M_c )</td>
</tr>
<tr>
<td>( I/M R^2 )</td>
</tr>
<tr>
<td>( \rho_0 ) mantle</td>
</tr>
<tr>
<td>( \rho_0 ) core</td>
</tr>
</tbody>
</table>
| \( R_c \), in km, is the core radius; \( M_c \), in wt.%, is the core mass; \( I/M R^2 \) is the dimensionless value of the moment of inertia; and \( \rho_0 \) is the zero-pressure volume mass. Error bars include results for crustal thickness ranging from 20 to 50 km, crustal density from 3.0 to 3.2 and \( K_0 \) mantle from 130 to 170 GPa. Longhi’s model is based on the Wänke and Dreibus’ chemical model.

(5)

(where \( \rho_{mantle} \), the volumic mass, depends upon \( \rho_0 \) through Eq. (5)). For the mantle, different values for \( K_0 \) have been tested, ranging from 130 GPa to 170 GPa (a domain which roughly covers the \( K_0 \) values for main silicate minerals: olivines, pyroxenes and garnets). In this range, results are not significantly affected by the value of the bulk modulus, \( K_0 \); those presented in Fig. 2 correspond to \( K_0 = 150 \) GPa, \( \alpha_0 = 2.6 \times 10^{-5} \) \( K^{-1} \), \( K'_0 = 5 \). In the thermal pressure term, \( \alpha K_{mantle} \) is considered constant and equal to \( \alpha_0 K_0 \), as attested by high pressure–high temperature experiments for a number of minerals because temperatures within terrestrial planets largely exceed the Debye temperature (Anderson et al., 1992). For the core, \( K_{FeS} = 170 \) GPa, \( K_{Fe} = 54 \) GPa, \( \alpha_{Fe} = 1.3 \times 10^{-5} \) \( K^{-1} \), \( \alpha_{FeS} = 7 \times 10^{-3} \) \( K^{-1} \), \( \rho_{Fe} = 7000 \) kg m\(^{-3}\) and \( \rho_{FeS} = 49400 \) kg m\(^{-3}\) (Fei et al., 1997).

The temperature profile corresponds to adiabatic gradients in the mantle and in the core, \( T/T_0 = (\rho/\rho_0)^{\gamma} \), by integration of the thermodynamic definition of the Grüneisen parameter, \( \gamma = (\partial T/\partial \rho)_S \), assuming \( \gamma \) constant (1.1 in the mantle and 1.4 in the core) and taking \( T = 1650 \) K as an upper estimate at the base of the lithosphere (Schubert and Spohn, 1990; Mocquet et al., 1996; Fei et al., 1997). In the conductive layers (lithosphere and core mantle boundary), temperature increases linearly with a 100 K thermal boundary layer at the core–mantle boundary (Fig. 2A).

The two chemical models lead to different core size and zero-pressure mantle density (Table 4 and
Fig. 2. Temperature, pressure and density profiles resulting from the present martian compositional models. The heavy line corresponds to the EH45:H55 composition and implies a value of 0.361 for the dimensionless value of the moment of inertia, $I/r_M R^2$, 3.45 for the zero-pressure mantle density, $r_{mantle}$; the dashed line corresponds to the EH70:H30 composition and implies only 0.352 for $I/r_M R^2$ and 3.30 for $r_{mantle}$. The value of 0.361 for $I/r_M R^2$ is increased to 0.363 by changing the thickness of the crust ($\rho_{crust} = 3000$ kg m$^{-3}$) from 40 km to 20 km, while taking $K_0 = 130$ GPa for the mantle. This value is similar to the lowermost bound recently reported (Folkner et al., 1997).

5. Discussion

This computed physical structure of Mars enables us to tightly constrain the Martian mineralogy while checking its consistency with the chemical model. We will therefore begin the discussion by comparing the present chemical model with previous ones.

5.1. Chemical composition

Among the major elements, only Na and S are volatile elements and merit a few additional comments. The best estimate for Na in Mars mantle is certainly derived from the SNC composition, as done by Dreibus and Wänke (1985). Our estimate is only a maximum value assuming no loss upon accretion. The case of S is more critical. Again, our result gives a maximum value for S in the core; because of S volatility, the actual amount may be smaller, but we have no decisive observation for that. However, we can test the effect of S loss on the physical properties of Mars. The amount of S in the core changes its mass: decreasing S by a factor of 2 will change the mass of the core by about 7% and will lower the moment of inertia 0.358 as the density of the core will increase if it contains significantly less sulfur. Comparison with the previous estimate of Dreibus and Wänke (1985) is interesting. They calculate the amount of S in the core from the abundance of chalcophile elements in the SNC (Co, Ni, W, Zn and Cu). They assume that their low concentrations result from sulfide segregation to the core from a chondritic composition. Using the sulfur/silicate partition coefficients, they obtain a core representing 22% of the planetary mass with 14% S (vs. 23% with 16% S in our model).

A review of previous compositional models has already been presented (Longhi et al., 1992), but the comparison with the models of Wänke and Dreibus (1988) and Longhi et al. (1992) based mostly on SNC composition is worth being discussed here in detail. First, the mass fraction of core and mantle in their model was adjusted to fit the moment of inertia and density. Here, the core to mantle ratio is given
by the chemical composition alone. Our core/mantle ratio 23:77 is very close to theirs of 22:78, though the methods are independent and strongly different. The agreement is within the uncertainties of both approaches. The similarity of the core composition results mostly from the nearly similar S content 16.2% here instead of 14.2% (see discussion above). The mantle composition is worth being discussed in greater details. Most of the results of Dreibus and Wänke (1985) were dictated by using the geochemical similarities of a number of elements. Among the major elements, Mg, Al, Si, Ca and Ti are supposed to occur in CI abundance, as we know from Earth and Moon, this assumption holds within an error of about 30%. Mg and Si exhibit the major differences, since among chondrites, CI have the largest Mg/Si and the values given by Dreibus and Wänke are not constrained by SNC data. The same holds for Al and Ca. Their value of MnO is CI chondritic (0.46%) compared to our value of 0.40%. As far as MnO is concerned, it is a very minor difference. More important, however, is the determination of FeO from the nearly constant value of FeO/MnO = 39.1 ± 1.5 as measured in SNC. Our FeO/MnO is 38.0, i.e., within the range of SNC which corresponds to FeO = 15.4 instead of 17.9. There is a very good agreement concerning the FeO/MnO ratio, even though Dreibus and Wänke FeO estimate is quite uncertain because of the limited accuracy on the starting value for MnO. Notice that our FeO/MgO = 0.55 is quite close to theirs (0.59) especially when compared to the terrestrial mantle (0.213) (McDonough and Sun, 1995). This point is important; the terrestrial mantle differs drastically from the Martian mantle which illustrates that the similarity between our FeO/MgO and that derived from the study of SNC cannot be coincidental. Rather than the Mg and Fe abundance in the chondritic precursor, the important parameter is the abundance of oxygen (Fe/FeO ratio) which is highly variable among different chondritic classes, a point explicitly considered in our model for the first time.

Minor elements Cr, Co and Ni are determined from their relationship with Mg or (MgO + FeO). Our estimate for Cr2O3 is 0.66 compared to their CI value of 0.76. For Co and Ni, our model provides no value. If we use the same correlations as Dreibus and Wänke and using our Mg and Fe results, we obtain 60 and 350 ppm for Co and Ni, respectively.

5.2. Mineralogical composition

The major differences between our model and that of Dreibus and Wänke can be clearly illustrated when calculating an upper mantle mineralogical norm (Table 5). In our model, orthopyroxene is by far the dominant phase, whereas in Dreibus and Wänke, olivine and orthopyroxene are in ‘terrestrial’ proportions. The dominance of clinopyroxene and majoritic garnet was recently confirmed by multianvil experiments conducted on the EH70:H30 composition (Guén and Boivin, personal communication). It should be noticed that Nakhla, Lafayette, Governor Valladares and ALH84 are pyroxenites, while this kind of rock is very rare on earth. It is therefore anticipated that fractionation processes on Mars differed significantly compared to the Earth with pyroxene and garnet at depth, playing a major role as liquidus phases instead of olivine. Indeed, our compositions were used as input data for the petrologic software MELTS (Ghiorso and Sack, 1995) and with both of them, pyroxene is given as the liquidus phase.

Table 5
Simple mineralogical mode (in wt.% and at ambient conditions) for the terrestrial primitive mantle and Martian mantle models

<table>
<thead>
<tr>
<th>Mineral name and formula</th>
<th>EH45±H55</th>
<th>EH70±H30</th>
<th>Dreibus and Wänke, 1985</th>
<th>Primitive mantle</th>
</tr>
</thead>
<tbody>
<tr>
<td>Olivine (SiO₃)·(MgO,FeO)</td>
<td>26</td>
<td>4</td>
<td>45</td>
<td>50</td>
</tr>
<tr>
<td>Orthopyroxene (SiO₃)·(MgO,FeO)</td>
<td>60</td>
<td>77</td>
<td>40</td>
<td>28</td>
</tr>
<tr>
<td>Ca-Pyroxene (SiO₃)·Ca(MgO,FeO)</td>
<td>3</td>
<td>2</td>
<td>3</td>
<td>4</td>
</tr>
<tr>
<td>Garnet (SiO₃)·Al₂O₃·CaO(MgO,FeO)</td>
<td>11</td>
<td>13</td>
<td>12</td>
<td>17</td>
</tr>
<tr>
<td>x = (Mg#)</td>
<td>0.72</td>
<td>0.80</td>
<td>0.75</td>
<td>0.91</td>
</tr>
</tbody>
</table>

The mantle rock densities can then be calculated according to the Mg/(Fe + Mg) ratio (Mg#) of each model using empirical mineral densities as a function of their iron content (Duffy and Anderson, 1989). The compatibility of such values with the ones derived from our physical model enables us to discriminate between the two possible compositional models. Indeed, the mineralogical mode corresponding to the EH45:H55 chemistry implies a density of 3450.0–3500.0 (zero-pressure value) which agrees very well with the 3450.0 value (Table 4). On the contrary, the mantle composition for the EH70:H30 model and its Mg# (0.80) leads to a mineralogical paragenesis with a zero-pressure density of 3380 ± 20, which is significantly higher than the one derived from the physical model of 3300. This discrepancy is a strong argument to reject this EH70:H30 compositional model. Moreover, the FeO content for this model (11.35%) could be too low to be compatible with the 20% to 27% SNC’s FeO content (Longhi and Pan, 1989) if one assumes that they are produced by a standard melting event.

Pyroxene therefore seems to dominate the Martian mantle mineralogy. The presence of garnet is also expected, especially at depth where pyroxene transforms to garnet with increasing pressure and temperature. Such a transformation, which occurs gradually between 16 and 18 GPa, will lead to a smooth variation in the density profile instead of a sharp transition phase as would be the case in an olivine-rich mantle (Presnall, 1995); the 2% expected density variation will be spread over 200 km and will only marginally affect the calculated moment of inertia.

Very different mineralogical properties would also result from such a high pyroxene content and low Mg#. The viscosity for a pyroxenite is experimentally determined to be around 10 times lower than for a dunite in dry conditions (Kirby and Kronenberg, 1984; Karato and Wu, 1993). Moreover, the effect of Fe increases significantly the bulk modulus of majorite (high pressure garnet phase). Precise measurements of $P-V-T$ equations of state are therefore needed for Fe-rich pyroxenes and garnets for further investigation of the dynamics of Mars interiors.

Most of the results presented here should be testable soon with the next Mars mission to come, with the seismic experiments. This is especially true for the high pyroxene content we predict for the Martian mantle mineralogy, but also for its iron content.

6. Conclusion

A chemical model for Mars is proposed without relying on the SNC’s chemical compositions. However, these later can be used successfully to test the model validity in agreement with previous works (Dreibus and Wänke, 1985; Laul et al., 1986; Longhi et al., 1992).

The presented model should be testable soon with the future Martian missions in which seismic prospecting will be performed. In particular, these missions should check: (1) the absence of major seismic discontinuities in the mantle, (2) the core/mantle boundary depth and the core S-content, and (3) the mantle iron content and mineralogy. Also, the crust (and possibly lithosphere) thickness should be determined and will therefore help us in improving the precision of our model.

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