

# AN OXYGEN ISOTOPE MIXING MODEL FOR THE ACCRETION AND COMPOSITION OF ROCKY PLANETS

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**Abstract.** The oxygen isotope systematics in planetary and nebular matter are used to constrain the types of nebular material accreted to form a planet. The basic assumption of this model is that the mean oxygen isotopic composition of a planet is determined by the weighted mean oxygen isotopic composition of nebular matter accreted by the planet. Chondrites are taken as representatives of nebular matter. The chemical composition (which determines core size, mantle oxidation state, density, moment of inertia) of a planet results from the weighted mean compositions of the accreted nebular material, once the mass fractions of the different types of accreting matter are known. Here some results for Earth, Moon, Mars, and Vesta are discussed. The model implies that loss of volatile elements, such as alkalis and halogens, occurs during accretion and early planetary differentiation (*e. g.*, by catastrophic impacts). The possible depletion mechanisms of moderately volatile elements are discussed.

## 1. Introduction

The Earth, Moon, Mars, and Vesta are the only rocky planetary bodies where we currently have sufficient geophysical and geochemical information to test accretion models. The bulk densities are known for all terrestrial planets but the moment of inertia, which gives insight about the mass distribution within a planet, is only known for the Earth, Moon, and Mars. The chemical composition of bulk planetary silicates (mantle and crust) is known for the Earth and to some extent for the Moon. Information about the silicate composition of Mars and Vesta is primarily based on the general consensus that the shergottites, nakhlites, and chassignites (SNC) meteorites are from Mars and that the eucrites, howardites, and diogenites (EHD) meteorites are from Vesta. The discussion here is restricted to the Earth-Moon system, Mars, and Vesta because geophysical and geochemical data for Mercury and Venus are still insufficient for our purposes.

Many accretion models for the terrestrial planets call for compositionally different components to reproduce characteristic planetary properties (silicate oxidation state, core composition, volatile content, density, core size). The relative abundances of refractory lithophile elements in planetary silicates are comparable to those in chondritic meteorites suggesting a kinship between the relatively unaltered, primitive solar nebula material sampled by chondrites and the planets. Indeed, many accretion models for the Earth, Mars, or Vesta invoke chondritic matter in more or less modified form as building blocks (see Lodders and Fegley, 1997). However, each planetary object has a characteristic oxidation state, metal and sulfide content, volatile element content, and oxygen isotopic composition which are unlike those seen in chondrites. Thus, no single group of chondrites is representative for the matter accreted by a planet.



## 2. The Oxygen Isotope Mixing (OIM) Model

The oxygen isotopic composition of chondritic matter and of the planets can be used to find the bulk composition of planets. This oxygen isotope mixing (OIM) model was originally developed by Lodders (1991) who applied it to the Earth and Moon, Mars, and Vesta. The OIM model can also be applied to other objects (*e. g.*, planets, satellites, asteroids) where the oxygen isotopic systematics are known. The reasons for choosing oxygen are that oxygen is the most abundant element in rock, oxygen is generally the first or second (after iron) most abundant element in rocky planets, and each body has a characteristic oxygen isotopic composition. Figure 1 shows the oxygen isotopic systematics for the Earth and Moon, SNC and EHD meteorites, and chondrites. The oxygen isotopic compositions are reported in the  $\delta$ -notation, relative to standard mean ocean water (SMOW) in permil ( $\text{‰}$ ):

$$\begin{aligned}\delta^{18}\text{O} &= \left[ \left( \frac{{}^{18}\text{O}}{{}^{16}\text{O}} \right)_{\text{sample}} / \left( \frac{{}^{18}\text{O}}{{}^{16}\text{O}} \right)_{\text{SMOW}} - 1 \right] \times 1000 \\ \delta^{17}\text{O} &= \left[ \left( \frac{{}^{17}\text{O}}{{}^{16}\text{O}} \right)_{\text{sample}} / \left( \frac{{}^{17}\text{O}}{{}^{16}\text{O}} \right)_{\text{SMOW}} - 1 \right] \times 1000.\end{aligned}\quad (1)$$

None of the chondritic meteorites plot into the same space as the SNC or EHD meteorites, showing that neither the SNC nor the EHD meteorites have a genetic relationship to one single group of chondrites. The Earth and Moon share common oxygen isotopic compositions with the enstatite meteorites. However, the enstatite meteorites are much more reduced than terrestrial and lunar rocks demonstrating that the enstatite meteorites were not the sole matter accreted by the Earth and Moon.

The OIM model assumes that the oxygen isotopic composition of a planet was established by the weighted mean oxygen isotopic composition of the nebular material accreted to form the planet. This assumption allows us to find the proportions of different types of nebular matter accreted by the planet. Chondrites are chosen as representative samples of nebular matter. Once the proportions of the different types of accreting matter are known, the silicate portion and core size, the composition of the silicate portion and core, the silicate oxidation state, and the trace element distribution in the bulk planet can be calculated from the well known chemical compositions of chondrites (Lodders and Fegley, 1998).

The oxygen isotopic composition of the Earth and Moon, Mars, or Vesta are matched by chondritic components defining a triangular plot around each object in the three oxygen isotope plot. For example, the O-isotopic composition for Mars plots inside a triangle formed by CI, CV, and H-chondrites (Lodders and Fegley, 1997). In general, OIM model calculations show that planetary oxygen isotopic compositions are matched with CI, CV-, and H-chondrites. In some cases, a combination of other chondrites may also give physically meaningful results. The CI, CV-, and H-chondrites were used by (Lodders, 1991) and Lodders and Fegley (1997) because they can be regarded as endmembers of chondrite compositions: the CI-chondrites are the most water and volatile-rich; the CV-chondrites are the most refractory-rich carbonaceous chondrites, and the H-chondrites are the

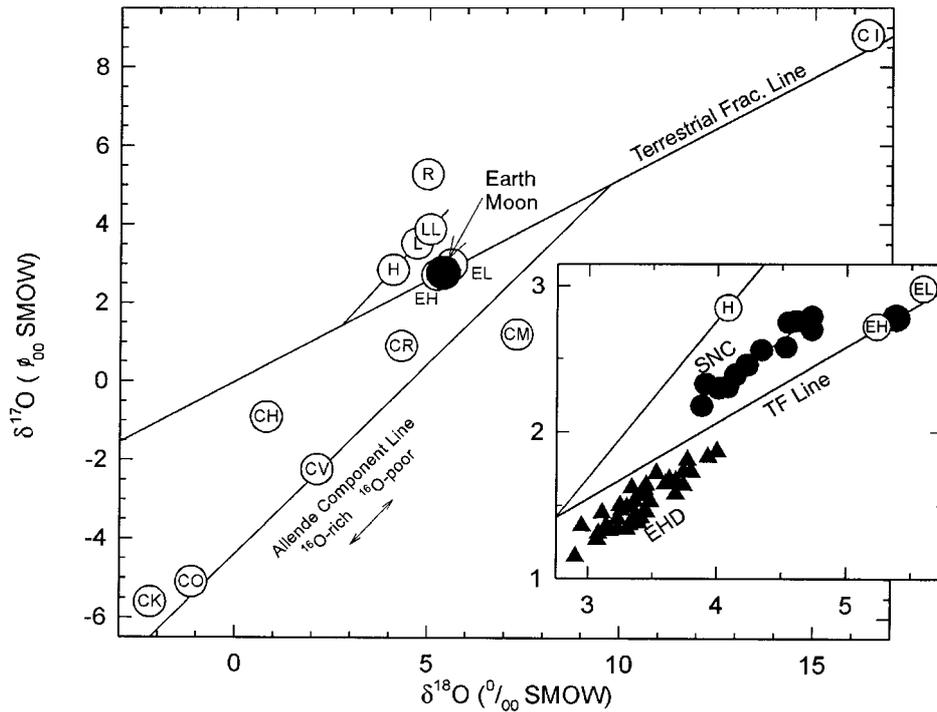


Figure 1. Oxygen isotope systematics for chondritic meteorites, the SNC and EHD meteorites, and the Earth and Moon. The insert shows an enlarged area for the differentiated meteorites (data from Clayton, 1993; Clayton *et al.*, 1991; Clayton and Mayeda, 1983; 1984; 1996; and references therein.)

most metal-rich ordinary chondrites. This combination yields good agreement with other independent model data for Mars and Vesta. These chondrite groups are also invoked in accretion modeling by other authors (see below).

The mass fractions ( $X_i$ ) of chondritic matter accreted by different planets is obtained by solving the three mass balance equations for  $^{16}\text{O}$ ,  $^{17}\text{O}$ , and  $^{18}\text{O}$  shown here for a mixture of CI, CV-, and H-chondrites:

$$\begin{aligned}
 ^{16}\text{O}_{\text{planet}} &= ^{16}\text{O}_{\text{CI}}X_{\text{CI}} + ^{16}\text{O}_{\text{CV}}X_{\text{CV}} + ^{16}\text{O}_{\text{H}}X_{\text{H}} \\
 ^{17}\text{O}_{\text{planet}} &= ^{17}\text{O}_{\text{CI}}X_{\text{CI}} + ^{17}\text{O}_{\text{CV}}X_{\text{CV}} + ^{17}\text{O}_{\text{H}}X_{\text{H}} \\
 ^{18}\text{O}_{\text{planet}} &= ^{18}\text{O}_{\text{CI}}X_{\text{CI}} + ^{18}\text{O}_{\text{CV}}X_{\text{CV}} + ^{18}\text{O}_{\text{H}}X_{\text{H}}
 \end{aligned}
 \tag{2}$$

and the condition that  $X_{\text{CI}} + X_{\text{CV}} + X_{\text{H}} = 1$ .

The abundances of  $^{16}\text{O}$ ,  $^{17}\text{O}$ , and  $^{18}\text{O}$  in each chondritic meteorite group and for the Earth, Moon, Mars and Vesta are from the oxygen isotope data reported as  $\delta$ -values by Clayton (1993), Clayton *et al.* (1991), and Clayton and Mayeda (1983; 1984; 1996; and references therein). The  $\delta$ -notation gives oxygen isotope ratios which cannot be used in mass-balance equations. It is necessary to convert data from the  $\delta$ -values to atomic percentages or absolute abundances for the calculations.

The uncertainties in the mass fractions from the mixing calculations arise from the standard deviation in the mean oxygen isotopic composition of the chondritic meteorites and the SNC or EHD meteorites. The uncertainties are mainly caused by the larger mean oxygen isotopic compositions of the CI and CV chondrites. As seen below, the ordinary chondritic matter supplies the largest fractions to Mars and Vesta, and enstatite chondritic and ordinary chondritic matter to Earth and Moon, so that the elemental abundances are primarily governed by these components. The uncertainties in the resulting bulk compositions for the planets also depend on the uncertainties of the chondritic elemental abundances, which are typically ten percent of the mean value for each element. More details of the model calculations, uncertainties, and applications to Mars are described by Lodders and Fegley (1997). The next sections briefly describe the results for Mars and give updated results for Earth and Moon, and Vesta (Lodders, 1991).

### 3. Calculated Planetary Properties

The mass balance calculations for the oxygen isotopes yield the mass fractions of the different chondritic components accreted by a planet. Then the bulk composition of the planet is calculated from the weighted mean composition of the different types of chondritic matter, using similar mass balance calculations for each element. However, it is important that the silicate, metal, sulfide and water content in the individual chondritic components are normalized to 100% before the mass balance calculations are done. This is particularly important for the carbonaceous chondrites which contain large amounts of water. We then compute the elemental abundances, the metal, sulfide, and water content in the accreting planet from mass balance equations

$$M_{\text{planet}} = M_{\text{CI}}X_{\text{CI}} + M_{\text{CV}}X_{\text{CV}} + M_{\text{H}}X_{\text{H}}, \quad (3)$$

where “M” stands for any elemental concentration or the metal, sulfide, or water content. The resulting metal and sulfide content in the planet then gives the relative core size. Water and carbon accreted from the chondritic components are likely to be involved in redox reactions with iron during accretion, so that total metal abundances will slightly increase (reduction of FeO by carbon) or decrease (oxidation of metal by water). This involves loss of H<sub>2</sub> and CO from the planet and the final planetary composition must be re-normalized to 100% to account for this small volatile loss (see Lodders and Fegley, 1997). The calculated bulk compositions for Mars and Vesta are shown in Tab. 1 together with results from other models. In addition, the silicate Fe# = [FeO/(FeO + MgO)] × 100 molar ratio, the mole fraction sulfur (X<sub>S</sub>) in the core, densities of the silicate portion, core and bulk planet, as well as the moment of inertia are listed. The densities (at STP) and dimensionless moments of inertia are recalculated here for all models listed in Tab. I to have a uniform base for comparison. The silicate densities for Mars use the high pressure mineralogy (see Lodders and Fegley, 1997) while low pressure mineralogy (CIPW norm) was used for Vesta.

TABLE I  
Model Compositions of Mars and Vesta

mass%	Mars			Vesta				
	OIM	Dela.	WD	OIM	BD	RD	DW	MHT
SiO <sub>2</sub>	36.03	35.9	34.8	36.37	34.36	38.3	36.22	35.0
MgO	23.58	22.9	23.6	23.75	22.25	25.13	24.70	24.8
Al <sub>2</sub> O <sub>3</sub>	2.29	2.1	2.3	2.29	2.14	2.84	2.56	2.2
CaO	1.87	1.75	1.9	1.92	1.78	2.16	2.02	1.8
TiO <sub>2</sub>	0.108	0.12	0.1	0.115	0.12	0.15	0.13	0.11
Na <sub>2</sub> O	0.78	0.79	0.4	0.753	0.67	0.75	0.086	0.04
K <sub>2</sub> O	0.088	0.09	0.03	0.084	0.08	0.08	0.007	0.004
P <sub>2</sub> O <sub>5</sub>	0.14	0.27	0.13	0.188	0.26	—	0.017	—
Cr <sub>2</sub> O <sub>3</sub>	0.538	0.51	0.6	0.517	0.50	0.5	0.61	—
MnO	0.290	0.30	0.4	0.282	0.28	0.30	0.370	—
FeO	13.66	11.2	14.0	13.74	13.72	26.89 <sup>b)</sup>	11.60	23.2
Fe <sub>2</sub> O <sub>3</sub>	0	0.37	0	0	0.28	—	0	—
C	<sup>a)</sup>	—	—	<sup>a)</sup>	0.63	—	—	—
H <sub>2</sub> O	<sup>a)</sup>	1.33	0.003	<sup>a)</sup>	—	—	—	—
% total silicates	79.37	77.63	78.3	80.01	77.07	97.1 <sup>b)</sup>	78.3	87.1
Fe-metal	12.68	14.2	11.52	11.79	11.10	—	19.92	11.2
Co	0.080	0.08	0.078	0.078	—	—	0.078	—
Ni	1.58	1.7	1.65	1.64	—	—	1.71	1.7
FeS	5.98	5.6	8.4	5.71	5.93	—	little	little
Fe <sub>3</sub> P	0.32	—	—	0.77	—	—	—	—
% total core	20.63	21.58	21.7	19.99	17.03	14	21.7	12.9
% total planet	100	99.21	100	100	94.1	97.1	100	100
Fe# silicates	0.24	0.21	0.25	0.24	0.25	—	0.21	0.34
FeO/MnO	47.1	38.4	35.0	48.7	49.0	—	31.4	—
total S in core, %	10.57	9.47	14.1	10.43	12.7	—	little	little
X <sub>S</sub> in core	0.17	0.15	0.23	0.17	0.20	—	—	—
$\rho$ silicates, g/cm <sup>3</sup>	3.50	3.46	3.52	3.35	3.38	—	3.36	3.53
$\rho$ core, g/cm <sup>3</sup>	7.27	7.38	7.04	7.18	7.00	—	8.15	8.15
$\rho$ total, g/cm <sup>3</sup>	3.92	3.94	3.95	3.75	3.73	—	3.85	3.66
C/(MR <sup>2</sup> )	0.367	0.363	0.367	0.367	0.370	—	0.360	0.387

OIM: Lodders (1991), Lodders and Fegley (1997), and this work.

Dela: Delaney (1994).

WD: Wänke and Dreibus (1988).

BD: Boesenberg and Delaney (1997).

RD: Righter and Drake (1997).

DW: Dreibus and Wänke (1980) and Dreibus *et al.* (1997).

MHT: Morgan *et al.* (1978).

<sup>a)</sup> Carbon and H<sub>2</sub>O from the chondritic components are considered in the mass-balance calculations and are assumed to escape as CO and H<sub>2</sub> after redox reactions involving Fe-metal and FeO (see text).

<sup>b)</sup> All Fe is reported as FeO. No distinction between iron in oxide, sulfide, and metal is given by Righter and Drake (1997).

### 3.1. MARS

The results for Mars described in detail by Lodders and Fegley (1997) are only briefly repeated here. The OIM model predicts that Mars accreted from a mixture of  $85.4 \pm 0.4\%$  H-,  $10.9 \pm 0.4\%$  CV-, and  $3.6 \pm 2.8\%$  CI-chondritic material. Not using O isotopes as a constraint, Anderson (1972) proposed a mixture of  $\sim 75\%$  ordinary- and  $\sim 25\%$  carbonaceous chondrites. Delaney (1994) derives a binary mixture of  $\sim 90\%$  H- and  $\sim 10\%$  CM chondrites to reproduce the martian O-isotopic signature.

The OIM model results can be compared to predictions from element correlations in SNC meteorites. For example, Wänke and Dreibus (1988) use the MnO/FeO ratio in SNC meteorites to find the fraction of total Fe in the silicates and core. However, to do so, they need to assume that elements such as Mg, Si, Al, Ca, Fe and Mn are present in CI-chondritic proportions. Wänke and Dreibus (1988) determine the abundances of volatile elements such as Na, or K by using volatile element over refractory element abundance ratios such as K/La or Na/Al. The assumption that La or Al are present in CI-chondritic abundances then again allows them to derive the abundances of Na or K from the observed element ratios. They do not employ element correlations to derive the amount of sulfur in the planet but instead, they assume that the depletion of sulfur is similar to that of elements of similar volatility such as Na and K. These model assumptions allow Wänke and Dreibus (1988) to determine the amount of metal and sulfur in the core. However, some important limitations of the element correlation method are discussed below.

Despite the different model assumptions, the calculated core mass of 20.6% and the Fe# of 24% from the OIM model agree with to the core size of 21.7% and Fe# of 25% from the SNC element correlation model by Wänke and Dreibus (1988). The results from the two component mixing calculations by Delaney (1994) are similar to those obtained by Lodders and Fegley (1997) and Wänke and Dreibus (1988) but yield lower FeO, Fe#,  $\text{Al}_2\text{O}_3$  and CaO in the silicates. The FeO abundance in Delaney's model would probably be higher if oxidation of iron metal by water (which Delaney gives as 1.33% for bulk Mars) would have been considered. The lower CaO and  $\text{Al}_2\text{O}_3$  abundances likely result from the use of CM chondrites in Delaney's mixing model while Lodders and Fegley use CV and CI chondrites. The abundances of Na and K from the OIM model by Lodders and Fegley (1997) and from the model by Delaney (1994) are higher than those obtained from element correlations of SNC meteorites. This important difference is addressed below.

The calculated bulk density ( $3.92 \text{ g cm}^{-3}$ ) and the moment of inertia ( $C/(\text{MR}^2) = 0.367$ ) from the OIM model agree with those observed for Mars ( $3.9335 \text{ g cm}^{-3}$ ,  $C/(\text{MR}^2) = 0.3662 \pm 0.0017$ , Folkner *et al.*, 1997) and are similar to the recalculated bulk density of  $3.95 \text{ g cm}^{-3}$  and  $C/(\text{MR}^2) = 0.367$  from the model by Wänke and Dreibus (1988). The data from Delaney's (1994) model give  $3.94 \text{ g cm}^{-3}$ , and  $C/(\text{MR}^2) = 0.363$ .

Due to limited space, I do not discuss the derivation of chalcophile and siderophile element abundances in the silicate portion of Mars from the OIM model here.

Their abundances are obtained from the calculated bulk composition and the use of (metal+sulfide)/silicate partition coefficients (Lodders and Fegley, 1997). The derived abundances compare well with independent results from element correlations in SNC meteorites.

### 3.2. VESTA

The OIM model predicts that Vesta accreted from  $77.9\pm 2.0\%$  H-,  $21.8\pm 4.8\%$  CV-, and  $0.3\pm 2.5\%$  CI-chondritic matter (Lodders, 1991, updated). Within uncertainties, a contribution of CI-chondrites to Vesta may not be required. The mass fractions obtained here for the H- and CV- component are similar to those obtained by other authors. Boesenberg and Delaney (1997) also use the oxygen isotopic composition to find mass percentages of 30% CM- and 70% H-chondrites for two component mixing to form Vesta, while Righter and Drake (1997) prefer a mixture of 30% CV- and 70% L-chondrites.

Contributions of CV- and H chondritic material to Vesta was noted previously, *e. g.*, a CV component by Jones (1984), the presence of H and CV or CO components by Morgan *et al.* (1978), and eucrite parent body similarities to H-chondrites by Dreibus and Wänke (1980). Shearer *et al.* (1997) find that a CV-like component is more suitable than CM, H, L, or LL chondrites because a component rich in incompatible elements is needed to satisfy incompatible element composition of magmas producing diogenites.

The model results for Vesta are also shown in Tab. I. The OIM model yields a core mass of 20% and a Fe# of 24% for Vesta. The calculated bulk density for Vesta from this model is  $3.75 \text{ g cm}^{-3}$ , which agrees with the observed density of  $3.8\pm 0.6 \text{ g cm}^{-3}$  reported by Thomas *et al.* (1997). Other calculated model densities also agree within the large error of the observed density (Tab. I).

The results by Boesenberg and Delaney (1997) yield a similar Fe# of 25% and a core mass of 18.1% (when re-normalized to 100%). Unfortunately, Boesenberg and Delaney (1997) and Righter and Drake (1997) used CM and CV chondrite abundances that do not total 100%. Thus, their calculated totals for Vesta are low (Tab. 1). Both groups also omit the water in carbonaceous chondrites in their calculations, which leads to some problems. Boesenberg and Delaney (1997) invoke dry CM chondritic matter but CM chondrites contain about 10% water (Lodders and Fegley, 1998). This means that  $\sim 20\%$  of the total oxygen in CM chondrites is in water. Neglecting this oxygen bound in water must lead to incorrect oxygen isotope mixing calculations. As mentioned above, water is also important because it can oxidize metal, which leads to a decrease of the core mass. However, Boesenberg and Delaney (1997) consider reduction of FeO by carbon because this favors a model FeO/MnO ratio closer to observed levels in eucrites.

The model results by Righter and Drake (1997) are difficult to compare since they report all iron as FeO, so that the Fe# or the metal content of the core cannot be easily computed. They give a core size of 14% but not its composition. They apparently neglected the presence of sulfur, carbon and water, and the fact that

chondritic components already contain metal and sulfide. Righter and Drake (1997) calculate possible core sizes (as mass percentages) from single chondritic sources. They obtain metallic cores of 20% from LL-, 18% from L-, and 15% from H-chondrites although LL chondrites are more oxidized and contain less metal and sulfide than H-chondrites, that should yield the bigger cores. Such core sizes can only be obtained by imposing an external redox buffer on the chondritic sources which would mean that Vesta (and, by analogy) any other planet did not accrete solely from chondritic components. External control of the redox-state must alter the bulk oxygen isotopic composition of the system. Thus, the oxygen isotopes from the accreting chondritic components cannot be used as a constraint to determine the types of accreting matter if an external gas phase (also containing oxygen) is imposed as a redox-buffer. This fact was not considered in the models by Righter and Drake (1997) or in the binary CM and H chondrite melting experiments under externally controlled oxygen fugacities by Boesenberg and Delaney (1997). In contrast, the OIM model only invokes the metal, sulfide, and FeO contents from the accreting components and the redox state is determined internally by Fe(metal)-FeO (silicate)-water-carbon equilibria.

Dreibus and Wänke (1980) and Dreibus *et al.* (1997) derive the silicate composition from howardite mixing relations and by addition of 43% of a pallasitic olivine component. Their core composition is obtained by assuming relative CI-chondritic abundances of Si, Fe, and Ni. Dreibus and Wänke obtain a core mass of 21.7%, which is sulfur free. Their FeO abundance and Fe# = 21% for Vesta's silicates are the lowest values among the Vesta models listed in Tab 1 and their FeO abundance is only slightly larger than their FeO abundance for Mars. The lower FeO abundance in Vesta's silicates allows for a bigger metallic core so that the absolute core mass (although sulfur free) is similar to that derived here from the OIM model.

The core mass of 12.9% derived by Morgan *et al.* (1978) is the lowest among the Vesta models. This is a consequence of a higher selected FeO content in the silicates. They obtain the bulk composition by considering nebular condensate components of varying volatility and taking a fixed MgO/FeO ratio as given by Stolper (1977) from eucrite melting experiments. Stolper's experiments suggest that the eucrite melt source region has an Fe# of 30 to 35%. None of the other models for Vesta listed in Tab. I give such a large Fe# for the bulk silicate portion. Other models for the silicate portion of Vesta (not shown in Tab. I) also based on Stolper's data are given by Jones (1984) and Consolmagno and Drake (1977). However, the adopted Fe# inferred from Stolper's experiments may only apply to the eucrite melt source region but may not be representative for the silicate portion as a whole. The bulk silicate composition must not only account for formation of eucrites during igneous differentiation but also for the diogenites with lower Fe# of ~27%.

One important difference between the various models is the composition of the core. The OIM model and the model by Boesenberg and Delaney (1997) are the only models which allow for a larger amount of sulfur in Vesta's core (~10%, Tab. I). Other models assume that sulfur is strongly depleted on the eucrite parent

body because sulfur is a volatile element like the alkali elements and the latter are severely depleted in EHD meteorites. However, this assumption may not be entirely justified. First, the volatility argument is based on condensation temperatures from a solar composition gas rich in H<sub>2</sub>, and the volatility of sulfur on a differentiating H<sub>2</sub>-gas free parent body is determined by the FeS-Fe buffer. Secondly, escape of sulfur from a parent body depends on the presence of sulfur-bearing gases produced from the initially accreted FeS. However, during differentiation FeS migrates together with metal to the core and may be taken out of the reaction network producing sulfur bearing gases. Thirdly, the most convincing argument that sulfur is present in larger quantities in the eucrite parent body is that the basaltic eucrites contain about  $0.16 \pm 0.08\%$  sulfur on average (Kitts and Lodders, 1998). This amount of sulfur is comparable to the average sulfur content of  $0.14 \pm 0.06\%$  in the Martian shergottites (Lodders, 1998), which are considered volatile rich. Thus, the volatility argument may fail for sulfur and the sulfur content of eucrites suggests that Vesta can have a larger amount of sulfur in its core.

One may argue that the matter accreting to Vesta was already depleted in volatile elements caused by fractionation during condensation from the solar nebula. The condensation temperature of sulfur is about 300 K lower than that of the alkalis. Thus, if the low abundance of alkali elements were established during incomplete condensation of the accreting components, sulfur should be depleted too. Then the eucrites should show very low abundances of alkalis and sulfur, which is not the case. The presence of sulfur and the low abundance of alkali elements in Vesta's silicates suggest that more selective loss of alkali elements may have occurred during accretion and early differentiation of Vesta; as discussed below.

### 3.3. EARTH AND MOON

The chemistry of Earth-Moon system is more difficult to model because chemistry during the giant impact forming the Moon is essentially unexplored. However, the OIM model gives the accreting components prior to any differentiation processes and thus gives a base for further modeling. The oxygen isotopic composition of the Earth-Moon system can be matched by a combination of ordinary and carbonaceous chondrite matter but the metal and sulfide content in these components cannot account for the large core size of the Earth. Therefore, enstatite chondritic matter is considered in the OIM model because enstatite meteorites have similar oxygen isotopic composition as the Earth and Moon, and the oxygen isotope constraint is not violated. However, in this case it is necessary to use the Earth's core size as an additional constraint to solve the set of mass-balance equations. The mass balance calculations then predict that the Earth accreted from about 70% EH-, 21% H-, 5% CV-, and 4% CI-chondritic matter. The bulk silicate resulting from this mixture has an Fe# of 13%, which is close to the Fe# of 11% observed for the Earth silicates. However, the detailed modeling for obtaining the silicate composition of the Earth and the Moon from the predicted composition after the giant impact will be reported elsewhere in the future.

#### 4. Moderately Volatile Elements (Alkalis, Halogens, Zn)

Like chondrite-based accretion models, the oxygen isotope mixing model predicts higher volatile element abundances than deduced from studies of planetary silicates. All planets are generally more depleted in volatiles relative to chondritic meteorites. The possible causes for the apparent volatile depletions for Mars, Vesta, and the Earth-Moon system are discussed next.

There are three possibilities for volatile depletion (1) volatile elements were depleted in the accreting matter; (2) these elements initially were present in chondritic abundance but volatilized during early stages of accretion and differentiation, or (3) these elements are now present in a reservoir not sampled or considered in modeling bulk planetary abundances.

The first possibility, that the planets accreted from matter already depleted in volatiles, is popular, but has some conceptual problems. The underlying concept for this assumption is that the volatile elements (*e.g.* alkalis, halogens, sulfur) condensed incompletely to proto-planetesimals in the solar nebula. These planetesimals then accreted to planets. This could occur by removing the planet forming matter from equilibrium with the solar gas prior to formation of chondritic matter, although proponents of this scenario have not described the physics of such a process yet.

Such a process implies two condensation scenarios from the solar nebula: incomplete condensation for all planet building matter and complete condensation for chondritic matter. Refractory elements are condensed before volatiles start condensing. If refractory-rich matter is removed from the solar nebula gas before volatiles condense, the refractory elements would be scavenged into planetary matter. In such a case, chondrites would be expected to be depleted in refractory elements, which is not the case. It would also imply that planetary matter is older than that of chondrites, for which no evidence is present; indeed Zhang (1998) argues for a younger age of the Earth than chondrites. In addition, volatile rich matter as seen in chondrites did form and the question is why this matter was not included in the population of planetesimals forming the planets.

It is also difficult to envision that chondritic matter condensed first and planetary matter later. The volatile elements are condensed when chondritic matter formed so any later forming planetesimals must also contain volatiles. In this case we may ask again why none of the volatile elements were incorporated in the later forming planetesimals accreting to the planets.

A plausible scenario is that the chondrite parent bodies (rich in volatiles) formed at similar times as the planetesimals forming the planets. If this is the case and one would want to make the planets out of volatile depleted matter, the problem is how depleted matter is selectively scavenged by the planets and the volatile rich matter is sorted into chondrites. It appears odd that larger volatile depletions seem to be inherent to the differentiated planets (Earth, Mars, Vesta) but not to the chondrite parent bodies. Thus, other mechanisms must be responsible for the observed volatile depletions in differentiated bodies.

The second possibility for volatile depletion in planets is that condensation led to chondritic matter which accumulated into larger planets, but then volatile loss occurred during early accretion and differentiation. This scenario may apply to Vesta, as described below.

The third possibility is that volatile elements in differentiated planets are in a reservoir not sampled or considered in establishing planetary bulk compositions. Here the K/La ratio serves as an example. Both elements have about the same compatibility so that their abundance ratio should not change during igneous differentiation. Because K is more volatile than La it is often argued that the K/La abundance ratio of volatile K to refractory La can be taken as an indicator of planetary volatile element depletion when compared to chondritic K/La ratios. However, as useful as the element ratio method may seem, it only represents the ratio in a particular planetary reservoir, not the entire planet. We can see this from the K and La abundances in the Earth using data summarized by Lodders and Fegley (1998). The terrestrial silicate mantle has a K/La ratio of about 360, which is clearly below the CI-chondritic value of 2340 indicating that the Earth is depleted in K relative to CI-chondrites and La. However, the K/La ratio from mantle rocks is not representative of that for the whole Earth. About 50% of all terrestrial K is located in the Earth's crust (neglecting the possibility that some K is sequestered in the core) while only 34% of all La is in the crust. The crustal K/La ratio of 710 is about twice that of the terrestrial mantle. After combining the crustal and mantle contributions to the K and La abundances, we find a K/La ratio for the bulk silicate Earth of 480. The point here is that neither the K/La ratio from mantle derived rocks nor that from the crust is representative for the terrestrial K/La value. As a corollary, the extent of volatile depletion can only be evaluated if all planetary reservoirs are considered.

#### 4.1. MARS

As described by Lodders and Fegley (1997), the abundances of moderately volatile elements for Mars from the OIM model are higher than those inferred from SNC meteorites. The calculated abundances are higher by factors of 1.4 (Zn), 2 (F, Cs), 4–5 (Na, K, Rb), and 6–8 (Cl, Br, I). As discussed in the previous section, there are three possibilities for this discrepancy.

The first possibility that volatiles never accreted to Mars is unlikely for the reasons given above. The second possibility is that volatile loss occurred during early accretion and differentiation. If the depletion is proportional to volatility, Zn and halogens with low condensation temperatures should be more depleted than Na and K with higher condensation temperatures. In particular, the relative abundances of Zn, Cl, Br, I should be about the same because they have similar volatilities. However, Zn is relatively more abundant than Cl, Br, or I in SNC meteorites and the model and observed Zn abundance agree quite well (factor of 1.4). Thus volatility loss may not be that important for Mars.

The higher model halogen and alkali element abundances versus the lower abundances derived from element correlations in SNC meteorites require another

explanation. It is well known that on the Earth, the alkalis and halogens (but not Zn) are concentrated in crustal rocks and/or in the oceans. If we only use terrestrial mantle rocks to derive the bulk silicate Earth composition, we would get the wrong answer for the alkalis and halogens because the major reservoirs (crust and ocean) are neglected. A similar situation may apply to Mars. The SNC meteorites are derived from Martian mantle rocks. If alkalis and halogens were leached out during hydrothermal activity and concentrated in the Martian crust and/or surface, the source region of SNC meteorites is depleted in these elements. Then the alkali and halogen abundances for bulk silicate Mars derived from the SNC meteorites must be too low because the crustal reservoir is not considered (see Sect. 4 above). The high abundance of chlorine (0.6%) on the Martian surface determined by the Viking probes supports the idea that hydrothermal alteration and leaching to the crust is a likely cause for the difference in predicted abundances from the oxygen isotope mixing model and abundances inferred from SNC meteorites.

Thus, it is plausible that Mars is not very strongly depleted in volatile elements relative to chondrites and that the apparently missing volatiles are in a Martian reservoir not sampled by the SNC meteorites.

#### 4.2. VESTA

The small planet Vesta is the most volatile element depleted body considered here. The overall volatile depletion increases from Mars (highest relative volatile abundances) to Earth to Moon to Vesta (lowest volatile content). There is no apparent correlation of volatile content with planetary heliocentric distance. If volatiles condensed at larger distances from the Sun in the solar nebula we may expect that Vesta accreted more volatile-rich matter than Mars.

The alkali-, Zn-, and halogen abundances are diagnostic for volatile depletion processes. In Vesta, Zn abundances are very low, indicating that volatile related fractionations caused the depletion of the halogens, alkalis and Zn. As mentioned above, the presence of sulfur in eucrites may argue against accretion of volatile depleted matter by Vesta. Thus, volatiles might have been initially present in the matter accreted by Vesta with abundances as calculated by the OIM model, but were subsequently lost during accretion and early differentiation. Smaller planets like Vesta are more susceptible to volatile element loss during accretion while more massive planets may only lose volatiles in early growth stages when the gravitational well can be overcome by escaping species. Another important mechanism for volatile loss is catastrophic impacts. Takeda (1997) notes that high speed impacts into the partly solidified crust of the eucrite parent body during early stages of accretion can lead to surface loss of Na, which is supported by his discovery of very calcic plagioclase (An<sub>98</sub>) in the A-881394 eucrite. Catastrophic impacts may also have caused the extensive volatile depletion on the Moon (and to some extent on the Earth).

It has been argued that volatile loss from planets is unlikely because this would have caused a fractionation of K isotopes, which is not observed (Humayun and Clayton, 1997). However, Esat (1996) argues that the conditions which can cause

Rayleigh fractionation of K isotopes are very restricted. Young *et al.* (1998) point out the importance of diffusive transport because Rayleigh fractionation requires an isotopically homogeneous reservoir. Isotope fractionations may not occur during evaporation when self diffusion in the solid or liquid reservoir is slower than the evaporation rate. Here we conclude that the absence of Rayleigh type fractionation of the K isotopes may rule out volatile evaporation from a well-mixed, homogeneous reservoir but it may not rule out volatile loss by other processes such as catastrophic impacts, where all matter becomes evaporated and conditions for Rayleigh type fractionation are not met.

## 5. Future Work

The oxygen isotope mixing model provides a quantitative approach to model bulk planetary compositions. Complete modeling for Mars shows that the model yields good agreement for core size, oxidation state and overall chemical composition when compared to available observations for Mars and independent models based on SNC meteorites (Lodders and Fegley, 1997). Early results from this model for Vesta and the Earth-Moon system are promising (Lodders, 1991). The modeling needs to be completed for the Earth-Moon system. More investigations are necessary to understand and quantify possible volatile depletion during early planetary formation.

The OIM model also may give information about the chondritic matter distribution in the early solar system because it reveals the proportions of primitive matter accreted to the rocky planets. These implications for solar nebula structure require further studies.

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