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Determining the possible building blocks of the Earth and Mars

Thomas H. BURBINE[1*] and Kevin M. O'BRIEN[2]

[1]Laboratory for Extraterrestrial Physics, NASA Goddard Space Flight Center, Greenbelt, Maryland 20771, USA
[2]11152 Wallingsford Road, Los Alamitos, California 90720, USA
[*]Corresponding author. E-mail: tburbine@lepvax.gsfc.nasa.gov

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Abstract–To determine the possible building blocks of the Earth and Mars, 225,792,840 possible combinations of the bulk oxygen isotopic and chemical compositions of 13 chondritic groups at 5% mass increments were examined. Only a very small percentage of the combinations match the oxygen isotopic composition, the assumed bulk FeO concentration, and the assumed Fe/Al weight ratio for the Earth. Since chondrites are enriched in silicon relative to estimates of the bulk Earth, none of the combinations fall near the terrestrial magmatic fractionation trend line in Mg/Si-Al/Si space. More combinations match the oxygen isotopic composition and the assumed bulk FeO concentration for Mars. These combinations fall near the trend for shergottite meteorites in Mg/Si-Al/Si space. One explanation for the difficulty in forming Earth out of known chondrites is that the Earth may be composed predominately of material that did not survive to the present day as meteorites. Another explanation could be that significant amounts of silicon are sequestered in the core and/or lower mantle of the Earth.

INTRODUCTION

All the planets in the solar system formed out of material from the solar nebula (e.g., O'Neill and Palme 1998). This material is believed to have started out as dust and then later coagulated to form planetesimals, which then accumulated into planets through collisions (e.g., Weidenschilling 2000). Chondrites appear to sample this primitive material and are currently subdivided into 13 groups, which have a variety of compositions (Table 1).

Detailed studies of the possible building blocks of the terrestrial planets require samples that can be used to estimate the bulk chemistry of the planets. This study will focus on trying to determine the possible building blocks of the Earth and Mars since samples of these two planets are available for laboratory analysis.

One of the most important constraints on possible building blocks for the Earth and Mars is their respective oxygen isotopic compositions. Rocks from the Earth and Moon fall along a line (called the terrestrial fractionation line) in diagrams of $\delta^{17}O$ (‰ relative to Standard Mean Ocean Water or SMOW) versus $\delta^{18}O$ (‰ relative to SMOW) (Fig. 1). Most chondrites do not plot on the terrestrial fractionation line except for the EH and EL chondrites, which have very similar oxygen isotopic values to the assumed value for the bulk Earth. Distances from this line are given as $\Delta^{17}O$ (‰) (= $\delta^{17}O - 0.52 \times \delta^{18}O$). Samples from the same parent body tend to fall on a mass-fractionation line, such as the terrestrial fractionation line, which has a slope of 0.52 (e.g., Clayton 1993). Martian meteorites have an average $\Delta^{17}O$ of $+0.32 \pm 0.01$‰ (Franchi et al. 1999), which provides evidence that the Earth and Mars formed from different materials.

Another constraint is the bulk FeO concentration of a planet, which can be derived for any body if the concentration of FeO for the basaltic surface and the partition coefficient of FeO between solid and liquid phases are known (Robinson and Taylor 2001). The composition of basalt is a function of the composition of its mantle source as modified by element partitioning during melting and crystallization (e.g., Treiman 2003). Robinson and Taylor (2001) estimate that the silicate Earth has an average FeO concentration of ~8 wt%, while silicate Mars has an average FeO concentration of ~18 wt%.

To determine what known chondritic material could possibly be the building blocks of the Earth and Mars, over 225 million possible combinations of different chondritic meteorites were examined to determine the best matches to Earth's and Mars' oxygen isotopic and bulk chemical compositions. Earlier studies (e.g., Lodders and Fegley 1997; Sanloup et al. 1999; Lodders 2000) have tried to match the Earth's and Mars' oxygen isotopic compositions and derive these planets' bulk chemical compositions, but these studies looked at far fewer possible combinations of meteoritic material. A similar type of "brute-force" analysis was done by Korotev (1997) to find the minimum number of lithologic

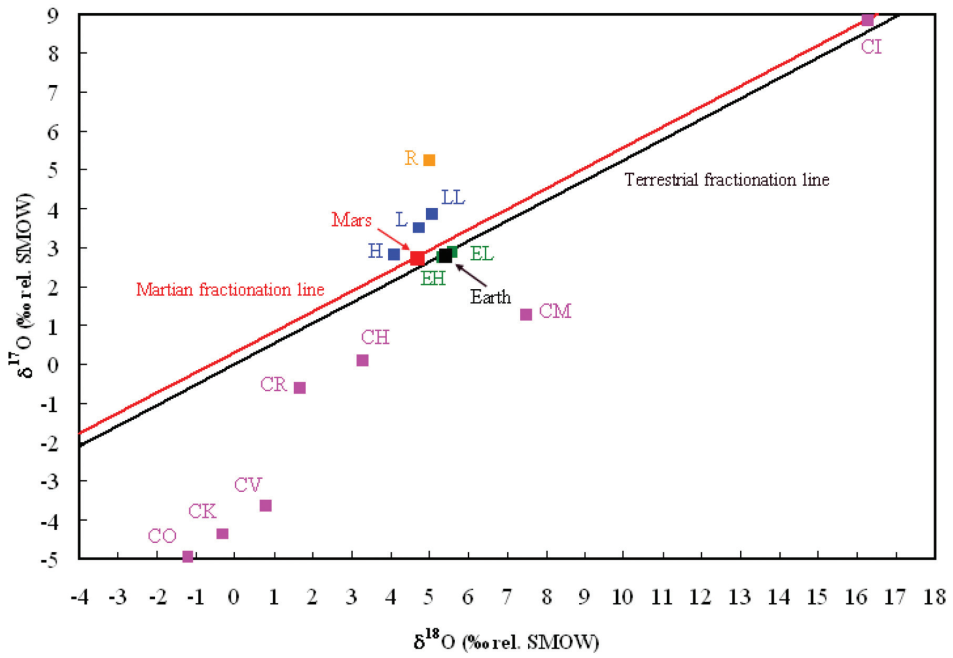

Fig. 1. Average oxygen isotopic values relative to Standard Mean Ocean Water (SMOW) are plotted for the chondritic meteorite groups (Table 4), the Earth (Eiler et al. 2000; Eiler 2001), and Mars (Franchi et al. 1999). The terrestrial fractionation and the martian fractionation lines are also plotted.

Table 1. Chondritic meteorite types.

Chondrites	Composition[a]
Carbonaceous	
CH	Pyroxene, metallic iron, olivine
CI	Phyllosilicates, magnetite
CK	Olivine, calcium-aluminum inclusions (CAIs)
CM	Phyllosilicates, tochilinite, olivine
CO	Olivine, pyroxene, CAIs, metallic iron
CR	Phyllosilicates, pyroxene, olivine, metallic iron
CV	Olivine, pyroxene, CAIs
Enstatite	
EH	Enstatite, metallic iron, sulfides, plagioclase, olivine
EL	Enstatite, metallic iron, sulfides, plagioclase
Ordinary	
H	Olivine, pyroxene, metallic iron, plagioclase, sulfides
L	Olivine, pyroxene, plagioclase, metallic iron, sulfides
LL	Olivine, pyroxene, plagioclase, metallic iron, sulfides
R	Olivine, pyroxene, plagioclase, sulfides

[a] Minerals or components are listed, approximately, in decreasing order of average abundance.

units that, when mixed, would account for the composition of the Apollo 16 regolith. Korotev (1997) looked at over 3 million combinations.

COMPOSITIONAL INFORMATION FOR EARTH AND MARS

The solid Earth can be divided into the crust, the mantle, and the core. The crust is dominated by basalts on the ocean floor plus granites on the continents (e.g., Ehlers and Blatt 1982). Peridotites, which are ultramafic rocks composed predominately of olivine plus pyroxene, are samples of the upper mantle (e.g., Jagoutz et al. 1979; Sun 1982; McDonough and Sun 1995; McDonough and Rudnick 1998). The composition of peridotites can be used to estimate the composition of the primitive upper mantle (PUM). (PUM is the hypothetical composition of the mantle before extraction of continental and oceanic basaltic crusts but after the formation of the core.) The core and lower mantle of the Earth cannot be directly sampled. However, geophysical observations (e.g., Bina 1998) of seismic waves and electric fields at the surface can be coupled with laboratory studies (e.g., Mao and Hemley 1998) to constrain the mineralogy and chemistry throughout the Earth. There is considerable discussion on how compositionally similar the PUM is to the lower mantle (e.g., Anderson 1981; Arculus et al. 1990; Helffrich and Wood 2001; van Keken et al. 2002).

Seismic studies show that the basic subdivision of the Earth is an upper mantle (down to a depth of ~410 km), a transition zone (~410–660 km), and a lower mantle (~660–2890 km) (e.g., Brown and Mussett 1993; Lodders and Fegley 1998; Agee 1998). At the boundaries of these transition zones, there are steep density gradients, which are generally believed to be due to the large number of high pressure phase changes of silicate minerals.

One of the major questions concerning the bulk composition of the Earth involves what elements, besides iron and nickel, are the major constituents of the Earth's core. Seismic studies (e.g., Birch 1952, 1964) and laboratory work (e.g., Brown et al. 1984) indicate that the Earth's core contains ~10% by mass of an element (or elements) lighter than Fe or Ni. A number of different elements (e.g., H, C, Mg, Si, S) have been proposed as being this less-dense component (e.g., Jeanloz 1990; Poirier 1994). Kilburn and Wood (1997), Sherman (1997), and Gessman et al. (2001) have shown that temperature and pressure constraints could allow the core to contain concentrations of Si of 7–9 wt%.

Drake and Righter (2002) argue that the primary building blocks of the Earth are some type of "Earth chondrite" or "Earth achondrite" that is not currently found in our meteorite collections. Estimates of the weight ratios of the Mg/Si and Al/Si of the PUM (Fig. 2) are depleted in Si relative to the chondrites (Table 2). The trend for differentiated terrestrial rocks in Mg/Si and Al/Si falls approximately perpendicular to the trend for chondritic meteorites. The trend for terrestrial rocks (peridotites, komatiites, basalts) is due to magmatic fractionation (differentiation) (e.g., Jagoutz et al. 1979; Drake and Righter 2002), as the extraction of magma tends to increase the Mg/Si ratio and decrease the Al/Si ratio in the mantle material. Possible solutions to the "apparent" depletion of Si in the PUM include sequestering Si in the core and/or the lower mantle.

As with the Earth, Mars can also be separated into different units: a basaltic crust, an olivine-dominated mantle, and an iron-rich core. Approximately 30 meteorites (including paired samples) are believed to originate from the martian surface. Martian meteorites are usually called SNCs for the shergottite, nakhlite, and chassignite meteorites. A number of spacecraft missions (e.g., the Viking landers, Mars Pathfinder, 2001 Mars Odyssey, Spirit, Opportunity) have visited Mars and have provided insight into this planet's bulk chemical and physical characteristics.

Shergottites include mafic and ultramafic rocks the chemical compositions and mineralogies of which have been used to constrain the composition of the martian mantle (e.g., Treiman 2003). The shergottite meteorites are separated (Meyer 2003) into the basaltic shergottites, olivine-phyric shergottites, and lherzolite shergottites. The olivine-phyric and lherzolite shergottites have mineralogies similar to

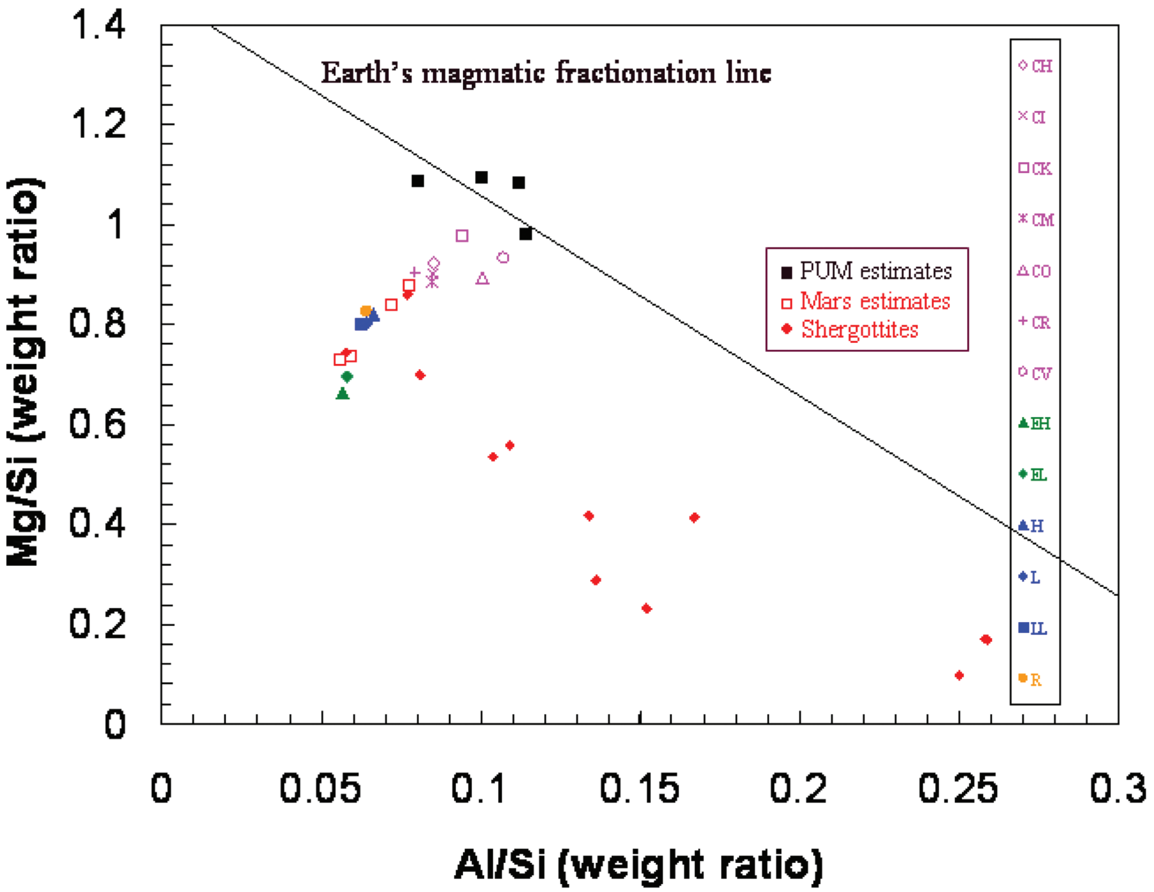

Fig. 2. Plot of Al/Si versus Mg/Si for the chondritic meteorites (average values), estimated values for the Earth's primitive upper mantle (PUM) (Al/Si = 0.08, Mg/Si = 1.09, Ringwood 1979; Al/Si =0.100, Mg/Si = 1.095, Jagoutz et al. 1979; Al/Si = 0.114, Mg/Si = 0.98, Palme and Nickel 1995; Al/Si = 0.112, Mg/Si = 1.086, McDonough and Sun 1995) (black squares), estimated values for Mars (Al/Si = 0.077, Mg/Si = 0.88, Wänke and Dreibus 1988; Al/Si = 0.072, Mg/Si = 0.84, Fegley and Lodders 1997; Al/Si = 0.059, Mg/Si = 0.88, Sanloup et al. 1999; Al/Si = 0.56, Mg/Si = 0.73, Mohapatra and Murty 2003) (open red squares), and values for shergottite meteorites (red diamonds) (Table 3). The line is the trend for terrestrial rocks, which is taken from Dreibus et al. (1998).

Table 2. Average Mg (wt%), Al (wt%), Si (wt%), Fe (wt%), Mg/Si (wt. ratio), Al/Si (wt. ratio), Fe/Si (wt. ratio), and Fe/Al (wt. ratio) for the chondritic meteorite groups.

	Mg[a] (wt%)	Al[a] (wt%)	Si[a] (wt%)	Fe[b] (wt%)	Mg/Si (wt. ratio)	Al/Si (wt. ratio)	Fe/Si (wt. ratio)	Fe/Al (wt. ratio)
CH	10.50[c]	0.97[c]	11.37[c]	45.09[c]	0.92	0.085	3.97	46.48
CI	10.32[d]	0.97[d]	11.43[d]	20.32[d]	0.90	0.085	1.78	20.95
CK	15.05[e]	1.45[e]	15.37[e]	25.24[e]	0.98	0.094	1.64	17.41
CM	12.21[f]	1.17[f]	13.79[f]	22.04[f]	0.89	0.085	1.60	18.84
CO	14.46[g]	1.63[g]	16.21[g]	24.91[g]	0.89	0.101	1.54	15.28
CR	14.05[h]	1.23[h]	15.51[h]	24.46[h]	0.91	0.079	1.58	19.89
CV	14.86[f]	1.71[f]	15.93[f]	23.64[f]	0.93	0.107	1.48	13.82
EH	11.32[i]	0.97[i]	17.08[i]	31.31[i]	0.66	0.057	1.83	32.28
EL	13.64[j]	1.14[j]	19.57[j]	25.88[j]	0.70	0.058	1.32	22.70
H	14.09[f]	1.14[f]	17.19[f]	27.56[f]	0.82	0.066	1.60	24.18
L	14.99[f]	1.20[f]	18.66[f]	22.04[f]	0.80	0.064	1.18	18.37
LL	15.33[f]	1.19[f]	19.13[f]	19.79[f]	0.80	0.062	1.03	16.63
R	13.78[k]	1.07[k]	16.67[k]	24.80[k]	0.83	0.064	1.49	23.18

[a] Concentrations of Mg, Al, and Si, respectively, were converted from the weight percents of MgO, Al_2O_3, and SiO_2. The only exceptions are the EH and EL chondrites where the Si (wt%) value includes Si in the metallic iron.

[b] The concentrations of Fe are calculated from the concentrations of FeO, Fe_2O_3, Fe (metal), and FeS.

[c] Calculated from data for NWA 470 (CH) from Ivanova et al. (Forthcoming).

[d] Calculated from data for Orgueil (CI) from Jarosewich (1990).

[e] Calculated from data for Karoonda (CK) from Mason and Wiik (1962b).

[f] Calculated from average value of falls from Jarosewich (1990).

[g] Calculated from average of data (Jarosewich 1990) for CO chondrites ALH A77003 and ALH 82101.

[h] Calculated from data for Renazzo (CR) from Mason and Wiik (1962a).

[i] Calculated from data for ALH A77295 (EH4) from Jarosewich (1990).

[j] Calculated from average of data (Jarosewich 1990) for EL6 chondrites Eagle and Pillistfer.

[k] Calculated from data for Rumuruti (R) from Jarosewich (unpublished data).

terrestrial peridotites (Meyer 2003). Mg/Si and Al/Si values for the shergottites are listed in Table 3. The Mg/Si-Al/Si trend for the shergottites falls far below the trend for terrestrial rocks, showing again that Mars is composed of different materials than those of the Earth.

Estimates of the bulk composition of Mars are usually derived from a cosmochemical model where the resulting composition fits the density distribution constraints imposed by the moment of inertia (e.g., Longhi et al. 1992). Knowledge of high pressure mineral phase transitions is also very important (e.g., Bertka and Fei 1998); however, since Mars is approximately one-eighth the mass of Earth, these high pressure phases will occur at deeper depths than they do on Earth (e.g., Halliday et al. 2001). These models (e.g., Longhi et al. 1992; Lodders and Fegley 1997) use elemental abundances in the SNCs and/or assumptions in the types of chondritic material that formed Mars. A number of estimates of the Mg/Si-Al/Si ratio for Mars are plotted in Fig. 2. The estimates overlap the values for chondrites since these models assume that they were formed on Mars from some mixture of chondritic materials.

METEORITE DATA

The average oxygen isotopic data for chondrites used in this study are listed in Table 4. Average $\Delta^{17}O$ (‰), $\delta^{17}O$ (‰), and $\delta^{18}O$ (‰) for each meteorite type were compiled primarily from the data of Clayton et al. (1991) and Clayton and Mayeda (1999). The exceptions are values for the EH and EL chondrites (Newton et al. 2000) and R chondrites (Schulze et al. 1994). The oxygen isotope values for the CM chondrites (Clayton and Mayeda 1999) do not include measurements of metamorphosed CM chondrites. The oxygen isotope values for the CV chondrites (Clayton and Mayeda 1999) include duplicate measurements of the meteorites Bali and Leoville since the oxygen isotopes of samples of these two meteorites are very different, apparently because these meteorites contain different abundances of refractory inclusions.

The chemical data was primarily compiled from the work of Jarosewich (1990) who did wet chemical analyses of ~200 bulk meteorite samples. Jarosewich (1990) measured the concentrations of all major elements and reported the results primarily as oxides. Other phases are presented as sulfides, SO_3, CO_2, elemental sulfur, H_2O, Fe (metal), and Fe_2O_3. For this study, concentrations of Si, Al, Mg, Fe, O, FeO, Fe_2O_3, Fe_3O_4, Fe (metal), C, and H_2O for each meteorite type (Table 2, Table 4, Table 5) were needed. Each of these elements and compounds were determined by Jarosewich (1990) or could be calculated from his data. The Appendix describes the decisions made in choosing the compositional data for each of the meteorite groups.

Compositional data for elements and compounds (Table 2, Table 4, Table 5) are listed with a number of

Table 3. Average Mg/Si and Al/Si weight ratios for shergottite meteorites. The classifications are based on those of Meyer (2003).

Martian meteorite	Classification	Mg/Si (wt. ratio)	Al/Si (wt. ratio)
ALH A77005	Lherzolitic shergottite	0.86[a]	0.077[a]
Dar al Gani 476/489	Basaltic shergottite	0.54[b]	0.104[b]
Dhofar 019	Olivine-phyric shergottite	0.41[c]	0.167[c]
EET A79001A	Olivine-phyric shergottite	0.42[a]	0.134[a]
EET A79001B	Basaltic shergottite	0.17[a]	0.258[a]
LEW 88516	Lherzolitic shergottite	0.70[a]	0.081[a]
Los Angeles 001/002	Basaltic shergottite	0.10[d]	0.250[d]
QUE 94201	Basaltic shergottite	0.17[a]	0.259[a]
Sayh al Uhaymir 005	Olivine-phyric shergottite	0.56[e]	0.109[e]
Shergotty	Basaltic shergottite	0.23[a]	0.152[a]
Yamato-793605	Lherzolitic shergottite	0.75[a]	0.058[a]
Zagami	Basaltic shergottite	0.29[a]	0.136[a]

[a] Calculated from data from Lodders (1998).

[b] Calculated from average of data for Dar al Gani 476 from Zipfel et al. (2000) and for Dar al Gani 489 from Folco et al. (2000).

[c] Calculated from data of Taylor et al. (2002).

[d] Calculated from data of Rubin et al. (2000).

[e] Calculated from data of Dreibus et al. (2000).

significant digits to attempt to get the most accurate results from the computations. However, there are many uncertainties in each calculated data value for each meteorite group. These uncertainties include relatively small analytical uncertainties in the measurements (Jarosewich 1990), assumptions on the distribution of elements into different oxides and compounds in the meteorites, corrections for adsorbed water and bound water due to terrestrial weathering, and possible problems with the use of one or very few meteorites as the source of data for a whole meteorite group. However, in the analysis of data from runs (~60 runs for the Earth and ~20 runs for Mars) using slightly different compositions for the meteorite groups, similar results and conclusions are achieved for each of the runs. Therefore, the results and conclusions appear to be relatively insensitive to these uncertainties.

METHODS

The program was written in the ANSI C++ programming language. The program generates all possible combinations of mass percentages for the chondritic groups at a user-specifiable increment (in this case 5%) that equal a total of 100%. The program was run on a 1.6 GHz Pentium IV computer. For the Earth, the run time of the program using 5% increments (225,792,840 possible combinations) was ~123 sec. A run time of ~14 days using 2% increments (over two trillion combinations) and a run time of ~76 yr using 1% increments (over 4 quadrillion combinations) is estimated. The use of a supercomputer, with appropriate program optimization, would shorten these times considerably.

If the calculated oxygen isotopic and bulk chemical values are within the tolerances (called a "match"), the results are stored, and several counters are incremented. These counters count the total number of matches and the total number of each percentage of each meteorite type that has been used in a match. The program also computes running averages and running variances of the percentages of each meteorite type that were used in matches. At the end of program execution, final averages and standard deviations (σ) (square roots of the variances) are available as the final value of these running values.

For each match for Earth, the distance to Earth's magmatic fractionation trend line in Mg/Si-Al/Si space is calculated. Dreibus et al. (1998) plotted this trend for terrestrial rocks, and this line is estimated to follow the equation $Mg/Si = -4.011 \times Al/Si + 1.460$. The program calculates the distance that any generated Mg/Si and Al/Si values calculated for a match falls from this line. The distances (d) to this line from any Mg/Si and Al/Si values calculated for a match were computed using the formula $d = \text{Absolute Value}[(ax + by + c)/(a^2 + b^2)^{1/2}]$, where a is the negative of the slope of the line, b is 1, c is the negative of the y-intercept of the line, x is the Al/Si value, and y is the Mg/Si value. It is possible to convert this distance to the percentage of Si that needs to be removed from a combination for it to intersect the line.

For each combination, the aggregate values of $\Delta^{17}O$ (‰), $\delta^{17}O$ (‰), and $\delta^{18}O$ (‰) are computed, along with weight percents of Si, Al, Mg, Fe, FeO, Fe_2O_3, Fe_3O_4, Fe (metal), C, and H_2O. Weight ratios of Mg/Si, Al/Si, Fe/Si, and Fe/Al are calculated as well. Elements and compounds are assumed to add linearly according to the mass percentage of each meteorite type in the combination except for the oxygen isotopic values. The $\Delta^{17}O$ (‰), $\delta^{17}O$ (‰), and $\delta^{18}O$ (‰) aggregate values are weighted by the oxygen concentration of each meteorite type.

Table 4. Average $\Delta^{17}O$ (‰), $\delta^{17}O$ (‰), $\delta^{18}O$ (‰), and O concentration (wt%) for the chondritic meteorite groups.

	$\Delta^{17}O$[a] (‰)	$\delta^{17}O$ (‰)	$\delta^{18}O$ (‰)	O[b] (wt%)
CH	−1.45	−0.60[c]	1.64[c]	24.74[d]
CI	+0.41	8.87[c]	16.26[c]	42.75[e]
CK	−4.16	−4.34[c]	−0.35[c]	37.03[f]
CM	−2.60	1.29[c]	7.49[c]	40.76[g]
CO	−4.28	−4.92[c]	−1.23[c]	35.70[h]
CR	−1.58	0.10[c]	3.24[c]	37.43[i]
CV	−4.01	−3.60[c]	0.78[c]	36.86[g]
EH	+0.01	2.76[j]	5.29[j]	28.84[k]
EL	+0.01	2.90[j]	5.56[j]	32.92[l]
H	+0.73	2.85[m]	4.08[m]	33.34[g]
L	+1.08	3.52[m]	4.70[m]	36.64[g]
LL	+1.26	3.88[m]	5.04[m]	38.09[g]
R	+2.70	5.28[n]	4.97[n]	36.36[o]

[a] The $\Delta^{17}O$ (‰) is calculated using the formula $\Delta^{17}O$ (‰) = $\delta^{17}O - 0.52 \times \delta^{18}O$.

[b] The concentrations of oxygen are calculated from measured oxides excluding adsorbed water plus bound water in the EH, EL, H, L, LL, and R chondrites.

[c] Average of whole rock analyses from Clayton and Mayeda (1999). The data for the CM chondrites does not contain measurements of metamorphosed CM chondrites.

[d] Calculated from data for NWA 470 (CH) from Ivanova et al. (Forthcoming).

[e] Calculated from data for Orgueil (CI) from Jarosewich (1990).

[f] Calculated from data for Karoonda (CK) from Mason and Wiik (1962b).

[g] Calculated from average value of falls from Jarosewich (1990).

[h] Calculated from average of data (Jarosewich 1990) for CO chondrites ALH A77003 and ALH 82101.

[i] Calculated from data for Renazzo (CR) from Mason and Wiik (1962a).

[j] Average values for enstatite chondrites from Newton et al. (2000).

[k] Calculated from data for ALH A77295 (EH4) from Jarosewich (1990).

[l] Calculated from average of data (Jarosewich 1990) for EL6 chondrites Eagle and Pillistfer.

[m] Average values for equilibrated ordinary chondrite falls from Clayton et al. (1991).

[n] Average value for R chondrites from Schulze et al. (1994).

[o] Calculated from data for Rumuruti (R) from Jarosewich (unpublished data).

Table 5. Average FeO, Fe_2O_3, Fe_3O_4, Fe (metal), C, and H_2O concentrations (wt%) for the chondritic meteorite groups.

	FeO (wt%)	Fe_2O_3 (wt%)	Fe_3O_4 (wt%)	Fe (metal) (wt%)	C (wt%)	H_2O (wt%)
CH	1.30[a]	–	–	42.94[a]	0.53[a]	2.96[a]
CI	1.31[b]	6.43[b]	11.71[b]	–	3.02[b]	11.64[b]
CK	29.00[c]	–	–	–	–	0.80[c]
CM	22.50[d]	–	–	0.14[d]	1.85[d]	8.89[d]
CO	20.70[e]	–	–	4.43[e]	0.17[e]	0.51[e]
CR	15.10[f]	–	–	10.52[f]	1.44[f]	5.56[f]
CV	26.90[d]	–	–	0.16[d]	0.43[d]	0.15[d]
EH	–	–	–	22.54[g]	0.43[g]	–
EL	–	–	–	20.39[h]	0.25[h]	–
H	10.30[d]	–	–	16.05[d]	0.11[d]	–
L	14.50[d]	–	–	7.07[d]	0.12[d]	–
LL	17.50[d]	–	–	2.46[d]	0.22[d]	–
R	27.40[i]	–	–	–	0.08[i]	–

[a] Calculated from data for NWA 470 (CH) from Ivanova et al. (Forthcoming).

[b] Calculated from data for Orgueil (CI) from Jarosewich (1990). The concentration of magnetite (Fe_3O_4) in Orgueil is assumed to be 11 wt% (Hyman and Rowe 1983) before adjusting for adsorbed water. The concentrations of FeO and Fe_2O_3 not in the magnetite is then determined.

[c] Calculated from data for Karoonda (CK) from Mason and Wiik (1962b).

[d] Calculated from average value of falls from Jarosewich (1990).

[e] Calculated from average of data (Jarosewich 1990) for CO chondrites ALH A77003 and ALH 82101.

[f] Calculated from data for Renazzo (CR) from Mason and Wiik (1962a).

[g] Calculated from data for ALH A77295 (EH4) from Jarosewich (1990).

[h] Calculated from average of data (Jarosewich 1990) for EL6 chondrites Eagle and Pillistfer.

[i] Calculated from data for Rumuruti (R) from Jarosewich (unpublished data).

For Earth and Mars, respectively, the values of the aggregate $\Delta^{17}O$ (‰), $\delta^{17}O$ (‰), $\delta^{18}O$ (‰) are first checked against respective tolerances. Values falling within all of these tolerances are then checked against some chemical constraints for each planet. For Earth, the bulk Fe/Al ratio and the bulk FeO (wt%) were used. For Mars, just the bulk FeO (wt%) was used.

REACTIONS

To calculate the "true" bulk FeO of the Earth and Mars, a number of redox reactions that should occur due to heating during planetary accretion need to be simulated. Three redox reactions (Equations 3–5) from Lodders and Fegley (1997) and two reactions (Equations 1 and 2) were used to account for Fe^{3+}. The ordering of the redox reactions is shown below:

$$6Fe_2O_3 + C \rightarrow 4Fe_3O_4 + CO_2 \quad (1)$$

$$2Fe_3O_4 + C \rightarrow 6FeO + CO_2 \quad (2)$$

$$C + 2H_2O \rightarrow CO_2 + 2H_2 \quad (3)$$

$$Fe\ (metal) + H_2O \rightarrow FeO + H_2 \quad (4)$$

$$FeO + 0.5C \rightarrow Fe\ (metal) + 0.5CO_2 \quad (5)$$

A sixth reaction from Lodders and Fegley (1997):

$$SO_3 + 1.5C + Fe\ (metal) \rightarrow 1.5CO_2 + FeS \quad (6)$$

never occurs in any of our model calculations for the Earth and Mars. CI, CM, and CR chondrites, which are assumed to be the only chondrites that contain SO_3 (Jarosewich 1990; Burgess et al. 1991), have H_2O concentrations (Table 5) large enough to react with all of the bulk carbon (Equation 3) in any possible combination of chondritic material containing these meteorites.

As with the work of Lodders and Fegley (1997), it was assumed that all excess carbon and H_2O are converted to CO_2 and H_2, which escape. The calculated FeO was then renormalized to account for these lost gases. Concentrations of Mg, Al, Si, and Fe were also renormalized to account for these lost gases.

It was found that the redox reactions do not significantly change the concentrations of FeO for the combinations. For the combinations that fall within the constraints set for Earth's oxygen isotopic and bulk chemical composition, the redox reactions, on average, decrease the weight percent of FeO by 1.13 wt%. For the combinations that fall within the constraints set for Mars' oxygen isotopic composition, the redox reactions, on average, increase the weight percent of FeO by 0.50 wt%.

There was no attempt to model a number of unconstrained factors that could possibly affect the calculated FeO concentrations. Carbon possibly could be one of the light elements in the core (e.g., Jeanloz 1990; Poirier 1994), which would reduce the amount of carbon available for the reactions. Atmospheric equilibria could affect H_2O/H_2 and CO_2/CO ratios (e.g., Arculus et al. 1990) and possibly not allow the reactions to go to completion. Neither of these effects is expected to significantly change the calculated concentrations of FeO for the resulting combinations since the change for the matching combinations for Earth using reactions that went to completion was only ~1 wt%.

OXYGEN ISOTOPIC CONSTRAINTS

For the Earth, the oxygen isotopic constraints used were $\Delta^{17}O$ (‰) = 0.00 ± 0.03, $\delta^{17}O$ (‰) = 2.81 ± 0.26, and $\delta^{18}O$ (‰) = 5.40 ± 0.50. The $\delta^{18}O$ (‰) value is an estimated value for the PUM (Eiler et al. 2000; Eiler 2001). For Mars, the oxygen isotopic constraints were $\Delta^{17}O$ (‰) = +0.32 ± 0.03, $\delta^{17}O$ (‰) = 2.75 ± 0.26, and $\delta^{18}O$ (‰) = 4.68 ± 0.50. The $\delta^{17}O$ (‰) and $\delta^{18}O$ (‰) values are the average values from the SNC meteorites from Franchi et al. (1999). The $\Delta^{17}O$ (‰) range was 0.03, which is approximately equal to a two-sigma uncertainty for the terrestrial and martian fractionation lines as calculated by Franchi et al. (1999). Ranges of ±0.50‰ for the $\delta^{18}O$ values and ±0.26‰ for the $\delta^{17}O$ values were assumed. The oxygen isotopic range chosen for Earth encompasses measured mantle samples (e.g., Mattey et al. 1994; Deines and Haggerty 2000).

BULK CHEMICAL CONSTRAINTS

For Earth, the Fe/Al (weight ratio) is assumed to be 20 ± 2. McDonough and Sun (1995), Allègre et al. (1995), and McDonough (2001) believe that one of the best-known compositional characteristics of the bulk Earth is the Fe/Al ratio, which they estimate as 20 ± 2. This Fe/Al ratio is derived by assuming that there is no Al in the core, the core has 85% Fe, the Fe concentration of the silicate Earth is known, and that the concentration of aluminum of the PUM from peridotite samples can be accurately estimated.

The distance in Mg/Si-Al/Si space that any chondritic mixture falls from the Earth's magmatic fractionation trend is also determined. Since chondrites have average Mg/Si and Al/Si ratios that are enriched in Si relative to this trend line (Fig. 2), no mixture of chondritic material will fall on or near this line. No elemental ratio constraints are used for Mars.

The FeO constraint used for the Earth is 5.4 ± 0.5 wt%. This number is determined by assuming that the PUM has a FeO concentration of 8.0 wt% (Robinson and Taylor 2001) and assuming that the core is only 32.5 wt% of the mass of the Earth. The range of ±0.5 wt% is chosen to reflect a 10% uncertainty range in the estimate of the bulk FeO concentration of Earth.

The FeO constraint used for Mars was 14.4 ± 1.4. This number is determined by assuming that the martian mantle has a concentration of FeO of 18.0 wt% (Robinson and Taylor 2001) and assuming that the core is 20% of the mass of Mars. The range of ±1.4 wt% is chosen to reflect a 10% uncertainty range in the estimate of the bulk FeO concentration of Mars.

RESULTS FOR EARTH

The run for Earth was done to find all of the possible combinations of meteorites that had Δ^{17}O (‰) of 0.00 ± 0.03, δ^{17}O (‰) = 2.81 ± 0.26, δ^{18}O (‰) = 5.40 ± 0.50, Fe/Al = 20 ± 2, and FeO (wt%) = 5.4 ± 0.5. For these constraints, there were 514 matches out of 225,792,840 possible combinations (0.00023%). Slight increases in the tolerances will increase the number of matches, but they do not change the fundamental results. The distribution of matching combinations for Earth for each percentage of the chondritic groups is shown in Table 6. The average compositions of the matches are shown in Table 7.

The matches, on average, contain ~55% EL chondritic material. All other meteorite types have average abundances in the matching combinations of 10% or less. The high proportion of EL chondritic material in the matches is due to the Earth's very similar oxygen isotopic composition (Fig. 1) and Fe/Al weight ratio to values for the EL chondrites and an FeO concentration that is only slightly more oxidized than EL chondrites. Wänke (1981) has proposed in his heterogeneous accretion model that the Earth was formed primarily out of highly reduced material like enstatite chondrites. Righter (2004) has also postulated that the Earth

Table 6. The distribution of matching combinations for Earth for each mass percentage of each chondritic meteorite type.[a]

%	CH	CI	CK	CM	CO	CR	CV	EH	EL	H	L	LL	R
0	500	339	409	178	369	318	351	222	0	185	117	114	383
5	14	175	105	303	132	117	155	161	0	149	95	128	101
10	0	0	0	33	13	48	8	95	0	84	110	108	25
15	0	0	0	0	0	24	0	32	0	62	81	88	5
20	0	0	0	0	0	7	0	4	0	26	58	51	0
25	0	0	0	0	0	0	0	0	0	7	39	24	0
30	0	0	0	0	0	0	0	0	0	1	14	1	0
35	0	0	0	0	0	0	0	0	0	0	0	0	0
40	0	0	0	0	0	0	0	0	13	0	0	0	0
45	0	0	0	0	0	0	0	0	56	0	0	0	0
50	0	0	0	0	0	0	0	0	95	0	0	0	0
55	0	0	0	0	0	0	0	0	143	0	0	0	0
60	0	0	0	0	0	0	0	0	143	0	0	0	0
65	0	0	0	0	0	0	0	0	52	0	0	0	0
70	0	0	0	0	0	0	0	0	12	0	0	0	0
75	0	0	0	0	0	0	0	0	0	0	0	0	0
80	0	0	0	0	0	0	0	0	0	0	0	0	0
85	0	0	0	0	0	0	0	0	0	0	0	0	0
90	0	0	0	0	0	0	0	0	0	0	0	0	0
95	0	0	0	0	0	0	0	0	0	0	0	0	0
100	0	0	0	0	0	0	0	0	0	0	0	0	0
Average	0.14	1.70	1.02	3.59	1.54	3.05	1.66	4.50	55.36	6.30	10.40	9.13	1.62
σ	0.81	2.37	2.02	2.88	2.57	4.66	2.52	4.81	6.59	6.44	8.39	7.29	3.06

[a] For each chondritic group, we list the number of matching combinations for percentages from 0 to 100% at 5% mass intervals. The average mass percentage for each group for a matching combination and the standard deviation (σ) of the averages are also listed. The sum of each column is 514, which is the number of matching combinations for Earth.

Table 7. Average and standard deviations (σ) for the weight percents of Si, Mg, Al, Fe, and FeO plus the average Mg/Si, Al/Si, Fe/Al, and Fe/Si weight ratios for matches for Earth and Mars.[a]

	Si (wt%)	Mg (wt%)	Al (wt%)	Fe (wt%)	FeO (wt%)	Mg/Si	Al/Si	Fe/Al	Fe/Si
Earth									
Average	18.76	14.01	1.18	25.31	5.42	0.75	0.063	21.46	1.35
σ	0.17	0.12	0.01	0.36	0.29	0.01	0.001	0.39	0.03
Mars									
Average	17.29	14.01	1.18	26.02	14.41	0.81	0.068	22.15	1.51
σ	0.64	0.46	0.03	1.94	0.80	0.01	0.001	2.25	0.17

[a] The given Mg/Si, Al/Si, Fe/Al, and Fe/Si weight ratios are the averages of the calculated weight ratios for matching combinations of Earth and Mars and not the ratios of the averages of the individual elements.

may have accreted out of reduced material similar in composition to enstatite chondrites, and the current FeO concentration of the Earth's mantle was produced by the reaction of metallic iron with water (Equation 4) brought by bodies such as comets, water-bearing planetesimals, or Kuiper Belt objects. Javoy (1995) and Lodders (2000) both proposed that the Earth formed primarily out of EH chondrites; however, EH chondrites are not a good match for the bulk composition of the Earth since the Fe/Al ratio (~32) of these meteorites is too high.

However, as expected, none of the 514 matches are good compositional analogues for the Earth if Mg/Si and Al/Si weight ratios are used. The positions of the matches in Mg/Si-Al/Si space fall (Fig. 3) far from the Earth's magmatic fractionation trend line. For the average Mg/Si and Al/Si values of the matches, 32% of the Si must be removed to reach the Earth's magmatic fractionation trend line, or the Mg and Al concentrations must be increased by a similar amount. Agee and Walker (1998) argue that the PUM was Mg-rich due to olivine flotation in a magma ocean, but such a mantle would not be expected to be enriched in Al..

The low number of matches is controlled by the assumed low FeO concentration of the Earth compared to most chondrites. There are 458,360 matches without the FeO constraint. Only the CH, EH, and EL chondrites have lower concentrations of oxidized iron (found in FeO, Fe_2O_3, and Fe_3O_4) than the Earth. Increasing the assumed FeO concentration of the Earth tends to increase the number of matches. For example, increasing the possible FeO concentration from 5.4 ± 0.5 wt% to 6.4 ± 0.5 wt% increases the number of matches from 514 to 1033. This increase in the FeO concentration also reduces the average abundance of EL chondritic material in the matches to 48%.

A number of processes could affect the FeO concentration of the Earth's mantle over time. Kasting et al. (1993) argue that the upper mantle was originally more reduced than it is today and has become more oxidized, possibly through the subduction of hydrated oceanic crust and the loss of hydrogen to space. A lower FeO concentration for the PUM will have less matching combinations and higher proportions of EL chondritic material in the matches. Francis et al. (1999) argues that the upper mantle has become depleted

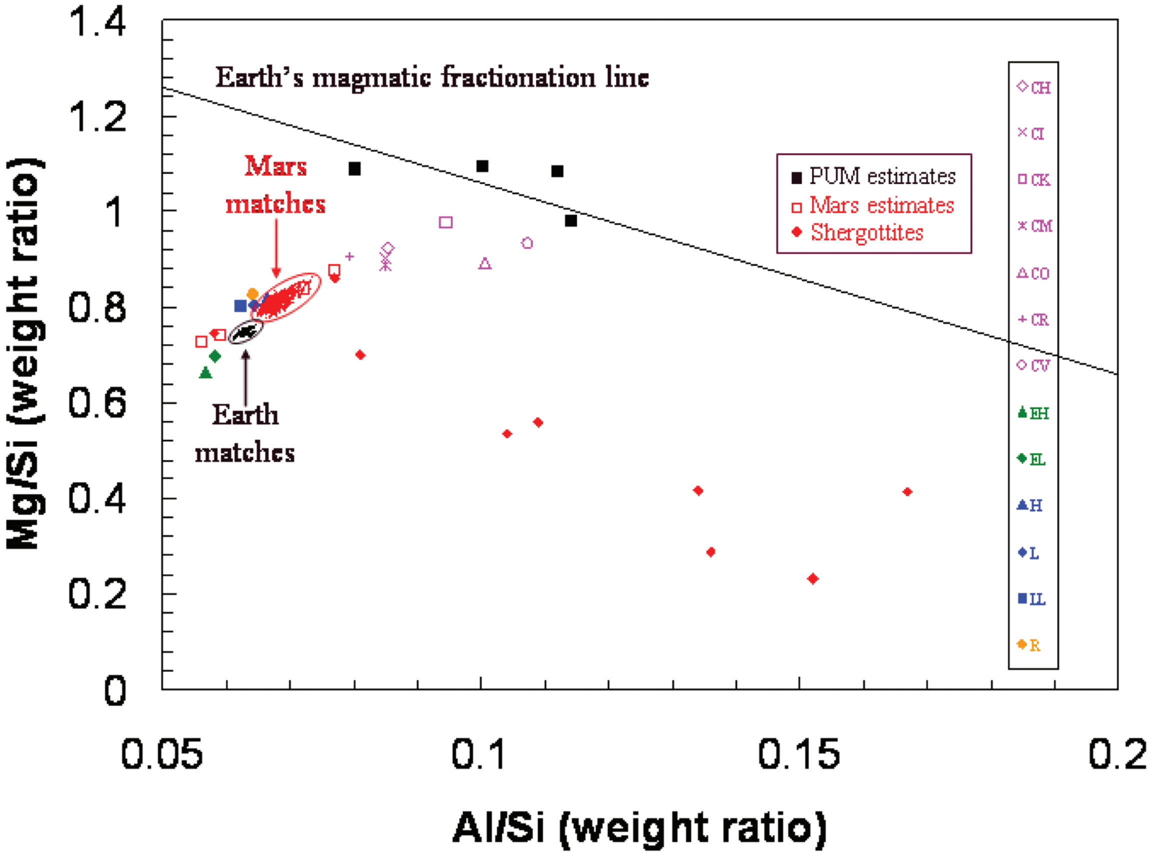

Fig. 3. Expanded plot of Al/Si versus Mg/Si from Fig. 2 for the chondritic meteorites (average values), estimated values for the PUM (black squares), estimated values for Mars (open red squares), and values for shergottite meteorites (red diamonds) (Table 3). The line is the trend for terrestrial rocks, which is taken from Dreibus et al. (1998). Ellipses are drawn around the estimates of the composition of the PUM and Mars. The black dots are the matches for the Earth, and the red dots are matches for Mars. Due to the high number of matches (over 165,000) for Mars, we only plot a distribution of 1000 points to show the range of matches for Mars.

in iron over time through the analysis of Archean (rocks older than 2.5 billion years) mantle samples, which are richer in iron than "modern" mantle samples. Francis et al. (1999) infer that the Archean mantle was up to 30% richer in iron. A higher FeO concentration for the PUM will result in more matching combinations and lower proportions of EL chondritic material in the matches.

If runs are done that try to find matches that just fall on the terrestrial fractionation line (no $\delta^{17}O$ and $\delta^{18}O$ constraints) (Burbine and O'Brien 2003), the matches tend to be dominated by CI chondrites since these meteorites have a $\Delta^{17}O$ value and a Fe/Al weight ratio similar to the Earth. However, such matches do not seem physically reasonable. For a planet dominated by CI chondritic material to have similar $\delta^{17}O$ and $\delta^{18}O$ values to the Earth and fall lower on or near the terrestrial fractionation line, a component with much higher $\delta^{17}O$ and $\delta^{18}O$ values must be removed.

Therefore, for Earth, it is possible to match estimates of Earth's oxygen isotopic composition and its bulk Fe/Al weight ratio and bulk weight percent of FeO using chondrites. However, as expected, the resulting bulk compositions calculated for the Earth plot far away from the Earth's magmatic fractionation trend line in Mg/Si-Al/Si space. Assuming the composition of the PUM falls on the Earth's fractionation trend line, it does not appear possible to make Earth out of known chondrites. The EL chondrites best match the Earth's oxygen isotopic composition, Fe/Al weight ratio, and FeO concentration.

RESULTS FOR MARS

The run for Mars was done to find all possible combinations of meteorites that had $\Delta^{17}O$ (‰) of 0.32 ± 0.03, $\delta^{17}O$ (‰) = 2.75 ± 0.26, $\delta^{18}O$ (‰) = 4.68 ± 0.50, and FeO (wt%) = 14.4 ± 1.4. For these constraints, there were 165,357 matches out of 225,792,840 possible combinations (0.073%). Slight increases in the tolerances will increase the number of matches, but they do not change the fundamental results. The distribution of matching combinations for Mars for each percentage of the chondritic groups is shown in Table 8. The average compositions of the matches are shown in Table 7 with the data for Earth.

The matches for Mars tend to be heavily weighted toward ordinary chondritic material. The H and LL chondrites, respectively, can make up 80%, and the L chondrites make up to 75% of a particular matching combination due to these meteorites falling just above Mars in the oxygen isotopic plot (Fig. 1) and having roughly similar weight percents of FeO. Compared to Earth, there is a much wider range among meteorite types for possible matches for Mars. Besides the ordinary chondrites, the CH, CR, EH, EL, and R chondrites are all part of matches with percentages of 40% or greater of these particular meteorite types.

The matches in Mg/Si-Al/Si space (Fig. 3) fall near the values for ordinary and R chondrites. Interestingly, the trend for the martian basaltic meteorites (shergottites) falls near where the matches for Mars plot. The matches appear

Table 8. The distribution of matching combinations for Mars for each mass percentage of each chondritic meteorite type.[a]

%	CH	CI	CK	CM	CO	CR	CV	EH	EL	H	L	LL	R
0	65619	82672	105220	84213	103414	70575	98804	51332	49310	42231	35455	31357	17892
5	42154	82473	46912	57309	45722	43154	46637	42505	41078	33003	29076	26727	24699
10	25570	212	12462	21586	14065	26015	16515	31427	31008	25394	23615	22710	29839
15	15270	0	763	2249	2152	14006	3340	20375	20894	19064	19156	19059	31055
20	8649	0	0	0	4	7442	61	11526	12487	14237	15146	15814	27374
25	4373	0	0	0	0	3020	0	5498	6548	10578	11917	12940	20086
30	2086	0	0	0	0	1059	0	2079	2855	7401	9151	10440	10941
35	1014	0	0	0	0	86	0	537	956	5164	6884	8263	3234
40	429	0	0	0	0	0	0	76	208	3495	5147	6301	237
45	122	0	0	0	0	0	0	2	13	2172	3671	4638	0
50	51	0	0	0	0	0	0	0	0	1284	2548	3205	0
55	20	0	0	0	0	0	0	0	0	731	1674	2055	0
60	0	0	0	0	0	0	0	0	0	356	1009	1166	0
65	0	0	0	0	0	0	0	0	0	158	548	502	0
70	0	0	0	0	0	0	0	0	0	68	256	167	0
75	0	0	0	0	0	0	0	0	0	19	84	13	0
80	0	0	0	0	0	0	0	0	0	2	20	0	0
85	0	0	0	0	0	0	0	0	0	0	0	0	0
90	0	0	0	0	0	0	0	0	0	0	0	0	0
95	0	0	0	0	0	0	0	0	0	0	0	0	0
100	0	0	0	0	0	0	0	0	0	0	0	0	0
Average	6.67	2.51	2.24	3.24	2.43	5.72	2.72	7.77	8.29	12.32	15.11	16.56	14.44
σ	7.83	2.51	3.26	3.78	3.53	6.67	3.79	7.58	8.03	12.46	14.50	14.90	9.13

[a] For each chondritic group, we list the number of matching combinations for percentages from 0 to 100% at 5% mass intervals. The average mass percentage for each group for a matching combination and the standard deviation (σ) of the averages are also listed. The sum of each column is 165357, which is the number of matching combinations for Mars.

consistent with the martian magmatic fractionation trend line as defined by the shergottite meteorites.

The average Fe/Si weight ratio (1.51) for the matches is similar to theoretical values for Mars computed by Bertka and Fei (1998) for core compositions of FeS and FeH, respectively. However, the standard deviation for the average Fe/Si weight ratio is 0.17, which means that the range of Fe/Si values for the matches encompasses all of the Fe/Si values calculated for Bertka and Frei (1998) for different core compositions. Our analysis does not allow us to make any determination of the light element in the core of Mars.

The results from this study are roughly consistent with those determined for Mars by Lodders and Fegley (1997). They found matches for Mars that were composed out of 85% H chondritic, 11% CV chondritic, and 4% CI chondritic material. Our matches for Mars overlap (Fig. 3) the Lodders and Fegley (1997) value. Our matches for Mars are enriched in silicon relative to the Wänke and Dreibus (1988) value since they used a significant amount of CI material in their model composition.

These matches for Mars are depleted in silicon relative to the values of Sanloup et al. (1999) and Mohapatra and Murty (2003). Sanloup et al. (1999) used oxygen isotopic values for the EH and H chondrites and geophysical models of Mars to fit Mars' bulk composition. Sanloup et al. (1999) found their best match for Mars used 55% H chondrite and 45% EH chondrite material. Mohapatra and Murty (2003) used nitrogen and oxygen isotopes in chondrites to compute possible building blocks of Mars. Mohapatra and Murty (2003) found matches for Mars that were composed out of approximately three-quarters enstatite chondritic material and one-quarter ordinary chondritic material. The primary reason for the discrepancy between our results and these two results is our use of the FeO constraint for Mars. Their mixtures would have concentrations of FeO (<6 wt%) that are too low to match the estimated value for Mars.

For Mars, it appears possible to match estimates of Mars' oxygen isotopic composition and bulk weight percent of FeO with known combinations of chondritic material. The matches for Mars also plot very near the trend for the shergottite meteorites in Mg/Si-Al/Si space, which seems to approximate the martian magmatic fractionation trend line.

DISCUSSION

Two solutions are usually proposed for the lack of mixtures of chondritic material that match the bulk chemical composition of the PUM. Either the Earth formed out of material that is not represented in meteorite collections and/or some of the Earth's Si is sequestered in the core and/or lower mantle. Adding in the proposed formation of the Moon from the impact of a "Mars-mass" body with the Earth (e.g., Canup and Asphaug 2001) would not noticeably help to make the bulk Earth out of chondrites. Since the Moon is only 1% of the mass of the Earth (Lodders and Fegley 1998) and estimates of the Moon's bulk Al/Si (0.099) and Mg/Si (1.02) ratios (O''Neill 1991a) fall only slightly closer to the chondrites than the estimates of the Earth's PUM, the bulk composition of the Earth-Moon system can not be readily matched by any mixture of chondritic material. O'Neill (1991b) has investigated the geochemical implications of a giant impact on the Earth to form the Moon and finds that the Earth's bulk composition before the impact still does not match any known chondritic material.

As suggested by Drake and Righter (2002), the Earth may have formed out of some unknown type of "Earth chondrite" or "achondrite" that matches the oxygen isotopic composition of the Earth and had Al/Si and Mg/Si ratios similar to the PUM. This type of material could have been totally consumed by the accreting Earth or exists today but has not fallen to Earth in recent history. Mars could have formed out of known chondritic material or an unknown type of "Mars chondrite" or "achondrite" that has similar oxygen isotopic values and bulk chemistries to known chondrites.

Wetherill (1994) argues that the idea of local feeding zones for the terrestrial planets is not valid and that all these planets received major contributions from bodies that were found between 0.5 and 2.5 AU and beyond. However, he believes that there is a correlation between the heliocentric distance of the planet and the average location of the bodies that make up the planet. More recent modeling (Chambers 2001) of the formation of the terrestrial planets from material between 0.3 and 2.0 AU also has found a compositional gradient among "produced" planets that preserves some memory of the initial distribution of the building blocks. The location of Mars closer to the asteroid belt could mean that Mars accumulated a higher proportion of the material that formed asteroids and that is now found in meteorite collections.

Some researchers (e.g., Gaffey 1990; Taylor and Norman 1990) believe that the Earth may have formed out of achondritic (differentiated) material. If the differentiated bodies have the same bulk composition as the chondritic bodies in our analysis, then our results will not be affected. Explosive volcanism, which has been proposed to eject volatiles and basaltic material from small differentiated bodies (e.g., Wilson and Keil 1991), could alter the bulk composition of these differentiated bodies. For the differentiated bodies to match the composition of the Earth, the lost material would have to be significantly enriched in Si relative to both Al and Mg in chondrites.

This hypothesis is tested by looking at the bulk compositions (Jarosewich 1990) of the basaltic meteorites known as the HEDs (howardites, eucrites, and diogenites), which are believed to be fragments of the crust of the differentiated asteroid 4 Vesta (e.g., Consolmagno and Drake 1977; Drake 2001). Among the HEDs, only the diogenites have both Al/Si (0.03) and Mg/Si (0.64) weight ratios lower than chondrites, indicating that they are enriched in Si relative to both Al and Mg in chondrites. The eucrites (Al/Si = 0.29, Mg/Si = 0.19) and howardites (Al/Si = 0.24, Mg/Si = 0.32) both have Mg/Si weight ratios lower than the chondrites, but their Al/Si weight ratios are much higher than chondrites due to the occurrence of high abundances of plagioclase in these meteorites. An Earth that formed out of chondritic material that differentiated and lost significant amounts of basaltic material similar in composition to diogenites (predominately orthopyroxene) could have a bulk composition similar to estimates of the Earth's PUM.

Sequestering Si in the Earth's core and/or lower mantle would also allow the Earth to be made out of known chondrites. All of the "missing" Si cannot be in the core since a "light" element or elements only make up 10 wt% of the core's mass. Assuming the bulk Earth contains 17.2 wt% Si (Allègre et al. 1995), our calculated 32% loss of Si from the PUM to intersect the Earth's magmatic trend line corresponds to 17.1 wt% Si in the core. Some but not all of the "missing" Si could be in the core since the core contains only ~10 wt% of a low-mass element. The rest of the Si would have to be sequestered in the lower mantle. Many researchers (e.g., Jeanloz and Williams 1998; Bina 1998) argue that current geophysical data appears consistent with a mantle with a peridotite bulk composition after taking into consideration increases in temperature and pressure with depth and mineralogical changes induced by pressure and that the data do not require significant enrichments of Si in the lower mantle. However, other researchers (e.g., Stixrude et al. 1992; Kellogg et al. 1999) argue that the geophysical data is consistent with a compositionally stratified mantle that is possibly enriched in Si in the lower mantle.

The most conclusive evidence for determining possible building blocks of the Earth would be the identification of a meteorite similar in composition to estimates of the bulk Earth or a mission to the Earth's core that can measure the composition of the lower mantle (Stevenson 2003). But in the near future, the most realistic method for gaining insights on the answer is to determine the possible building blocks of as many terrestrial planets as possible. Is Earth the only planet that is not made out of chondritic material? Or, is Mars the only planet that could be made out of chondritic material? Asteroid 4 Vesta can be analyzed in a similar fashion (Boesenberg and Delaney 1997; Burbine et al. 2003) as the Earth and Mars for determining possible building blocks if the HEDS are assumed to be fragments of asteroid 4 Vesta. Meteorites from Mercury and Venus appear dynamically possible (e.g., Gladman et al. 1996), but none have currently been identified. Planned spacecraft missions to Mercury (e.g., Solomon et al. 2001; Anselmi and Scoon 2001) may make it easier to identify mercurian meteorites on Earth.

CONCLUSIONS

This study's rigorous modeling using over 225 million possible combinations of chondritic material finds that it does not appear possible to match the Earth's assumed bulk chemical composition using known chondrites, but it is possible to match Mars' composition for our oxygen isotopic and bulk chemical constraints. A large number of combinations match Earth's oxygen isotopic composition. However, very few combinations match the Earth's assumed bulk FeO concentration, and none of the combinations fall close to the magmatic fractionation trend line in Mg/Si-Al/Si space for the Earth. Known chondrites are enriched in Si (or depleted in Mg and Al) relative to estimates of the PUM made from terrestrial rocks. Many matches can be found that match Mars' oxygen isotopic composition and bulk FeO concentration and that fall near the trend for shergottite meteorites in Mg/Si-Al/Si space.

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REFERENCES

Agee C. B. 1998. Phase transformations and seismic structure in the upper mantle and transition zone. In *Reviews in mineralogy, vol. 37, Ultrahigh pressure mineralogy: Physics and chemistry of the Earth's deep interior,* edited by Hemley R. J. Washington D.C.: Mineralogical Society of America. pp. 165–203.

Agee C. B. and Walker D. 1988. Mass balance and phase density constraints on early differentiation of chondritic mantle. *Earth and Planetary Science Letters* 90:144–156.

Allègre C. J., Poirier J. P., Humler E., and Hofmann A. W. 1995. The chemical composition of the Earth. *Earth and Planetary Science Letters* 134:515–526.

Anderson D. L. 1981. Hotspots, basalts, and the evolution of the mantle. *Science* 213:82–89.

Anselmi A. and Scoon G. E. N. 2001. BepiColombo, ESA's Mercury Cornerstone Mission. *Planetary and Space Science* 49:1409–1420.

Arculus R. J., Holmes R. D., Powell R., and Righter K. 1990. Metal-silicate equilibria and core formation. In *Origin of the Earth,* edited by Newsom H. E. and Jones J. H. New York: Oxford University Press. pp. 251–271.

Bertka C. M. and Fei Y. 1998. Implications of Mars Pathfinder data for the accretion history of the terrestrial planets. *Science* 281: 1838–1840.

Bina C. R. 1998. Lower mantle mineralogy and the geophysical perspective. In *Reviews in mineralogy, vol. 37, Ultrahigh pressure mineralogy: Physics and chemistry of the Earth's deep interior,* edited by Hemley R. J. Washington D.C.: Mineralogical Society of America. pp. 205–239.

Birch F. 1952. Elasticity and constitution of the Earth's interior. *Journal of Geophysical Research* 57:227–288.

Birch F. 1964. Density and composition of mantle and core. *Journal of Geophysical Research* 69:4377–4390.

Boesenberg J. S. and Delaney J. S. 1997. The composition of the basaltic achondrite planetoid. *Geochimica et Cosmochimica Acta* 61:3205–3225.

Brown G. C. and Mussett A. E. 1993. *The inaccessible Earth: An integrated view to its structure and composition.* London: Chapman & Hall. 276 p.

Brown J. M., Ahrens T. J., and Shampine D. L. 1984. Hugoniot data for pyrrhotite and the Earth's core. *Journal of Geophysical Research* 89:6041–6048.

Burbine T. H. and O'Brien K. M. 2003. Determining possible building blocks of the Earth (abstract #1193). Proceedings, 34th Lunar and Planetary Science Conference. CD-ROM.

Burbine T. H., O'Brien K. M., and Buchanan P. C. 2003. Chondritic building blocks of Vesta (abstract). *Geochimica et Cosmochimica Acta* 67:A49.

Burgess R., Wright I. P., and Pillinger C. T. 1991. Determination of sulphur-bearing components in C1 and C2 carbonaceous chondrites by stepped combustion. *Meteoritics* 26:55–64.

Canup R. M. and Asphaug E. 2001. Origin of the Moon in a giant impact near the end of the Earth's formation. *Nature* 412:708–712.

Chambers J. E. 2001. Making more terrestrial planets. *Icarus* 152: 205–224.

Clayton R. N. 1993. Oxygen isotopes in meteorites. *Annual Review of Earth and Planetary Sciences* 21:115–149.

Clayton R. N. and Mayeda T. K. 1999. Oxygen isotope studies of carbonaceous chondrites. *Geochimica et Cosmochimica Acta* 63: 2089–2104.

Clayton R. N., Mayeda T. K., Goswami J. N., and Olsen E. J. 1991. Oxygen isotope studies of ordinary chondrites. *Geochimica et Cosmochimica Acta* 55:2317–2337.

Consolmagno G. J. and Drake M. J. 1977. Composition and evolution of the eucrite parent body: Evidence from rare earth elements. *Geochimica et Cosmochimica Acta* 41:1271–1282.

Deines P. and Haggerty S. E. 2000. Small-scale oxygen isotope variations and petrochemistry of ultradeep (>300 km) and transition zone xenoliths. *Geochimica et Cosmochimica Acta* 64: 117–131.

Drake M. J. 2001. The eucrite/Vesta story. *Meteoritics & Planetary Science* 36:501–513.

Drake M. J. and Righter K. 2002. Determining the composition of the Earth. *Nature* 416:39–44.

Dreibus G., Ryabchikov I., Rieder R., Economou T., Brückner J., McSween H. Y., Jr., and Wänke H. 1998. Relationship between rocks and soil at the Pathfinder landing site and the martian meteorites (abstract #1348). 29th Lunar and Planetary Science Conference. CD-ROM.

Dreibus G., Spettel B., Haubold R., Jochum K. P., Palme H., Wolf D., and Zipfel J. 2000. Chemistry of a new shergottite: Sayh al Uhaymir (abstract). *Meteoritics & Planetary Science* 35:A49.

Ehlers E. G. and Blatt H. 1982. Petrology: Igneous, sedimentary, and metamorphic, New York: W. H. Freeman and Company. 732 p.

Eiler J. M. 2001. Oxygen isotope variations of basaltic lavas and upper mantle rocks. In *Reviews in Mineralogy and Geochemistry, vol. 43: Stable isotope geochemistry,* edited by Valley J. W. and Cole D. R. Washington D.C.: Mineralogical Society of America. pp. 319–364.

Eiler J. M., Schiano P., Kitchen N., and Stolper E. M. 2000. Oxygen-isotope evidence for recycled crust in the sources of mid-ocean-ridge basalts. *Nature* 403:530–534.

Folco L., Franchi I. A., D'Orazio M., Rocchi S., and Schultz L. 2000. A new martian meteorite from the Sahara: The shergottite Dar al Gani 489. *Meteoritics & Planetary Science* 35:827–839.

Franchi I. A., Wright I. P., Sexton A. S., and Pillinger C. T. 1999. The oxygen-isotopic composition of Earth and Mars. *Meteoritics & Planetary Science* 34:657–661.

Francis D., Ludden J., Johnstone R., and Davis W. 1999. Picrite evidence for more Fe in Archean mantle reservoirs. *Earth and Planetary Science Letters* 167:197–213.

Gaffey M. J. 1990. Thermal history of the asteroid belt: Implications for the accretion of the terrestrial planets. In *Origin of the Earth,* edited by Newsom H. E. and Jones J. H. New York: Oxford University Press. pp. 17–28.

Gessman C. K., Wood B. J., Rubie D. C., and Kilburn M. R. 2001. Solubility of silicon in liquid metal at high pressure: Implications for the composition of the Earth's core. *Earth and Planetary Science Letters* 184:367–376.

Gladman B. J., Burns J. A., Duncan M., Lee P., and Levison H. F. 1996. The exchange of impact ejecta between terrestrial planets. *Science* 271:1387–1392.

Halliday A. N., Wänke H., Birck J. -L., and Clayton R. N. 2001. The accretion, composition, and early differentiation of Mars. *Space Science Reviews* 96:197–230.

Helffrich G. R. and Wood B. J. 2001. The Earth's mantle. *Nature* 412: 501–507.

Hyman M. and Rowe M. W. 1982. Magnetite in carbonaceous chondrites (abstract). 13th Lunar and Planetary Science Conference. pp. 354–355.

Hyman M. and Rowe M. W. 1983. Magnetite in CI chondrites. Proceedings, 13th Lunar and Planetary Science Conference. *Journal of Geophysical Research* 88:A736–A740.

Ivanova M. A., Petaev M. I., Nazarov M. A., Taylor L. A., Barsukova L. D., Clayton R. N., Mayeda T. K., and Wood J. A. Forthcoming. The CH chondrite NWA 470: A mixture of fractional condensation products. *Meteoritics & Planetary Science.*

Jagoutz E., Palme H., Baddenhausen H., Blum K., Cendales M., Dreibus G., Spettel B., Lorenz V., and Wänke H. 1979. The abundances of major, minor, and trace elements in the Earth's mantle as derived from primitive ultramafic nodules. Proceedings, 10th Lunar Planetary Science Conference. pp. 2031–2050.

Jarosewich E. 1990. Chemical analyses of meteorites: A compilation of stony and iron meteorite analyses. *Meteoritics* 25:323–337.

Javoy M. 1995. The integral enstatite chondrite model of the Earth. *Geophysical Research Letters* 22:2219–2222.

Jeanloz R. 1990. The nature of the Earth's core. *Annual Review of Earth and Planetary Sciences* 18:357–386.

Jeanloz R. and Williams Q. 1998. The core-mantle boundary region. In *Reviews in Mineralogy, vol. 37, Ultrahigh pressure mineralogy: Physics and chemistry of the Earth's deep interior,* edited by Hemley R. J. Washington D.C.: Mineralogical Society of America. pp. 241–259.

Kasting J. F., Eggler D. H., and Raeburn S. P. 1993. Mantle redox evolution and the oxidation state of the Archean atmosphere. *Journal of Geology* 101:245–257.

Keil K. 1968. Mineralogical and chemical relationships among enstatite chondrites. *Journal of Geophysical Research* 73:6945–6976.

Kellogg L. H., Hager B. H., and van der Hilst R. D. 1999. Compositional stratification in the deep mantle. *Science* 283: 1881–1884.

Kilburn M. R. and Wood B. J. 1997. Metal-silicate partitioning and the incompatibility of S and Si during core formation. *Earth and Planetary Science Letters* 152:139–148.

Korotev R. L. 1997. Some things we can infer about the Moon from the composition of the Apollo 16 regolith. *Meteoritics & Planetary Science* 32:447–478.

Lodders K. 1998. A survey of SNC meteorite whole-rock compositions. *Meteoritics & Planetary Science* 33:A183–A190.

Lodders K. 2000. An oxygen isotope mixing model for the accretion and composition of rocky planets. *Space Science Reviews* 92: 341–354.

Lodders K. and Fegley B., Jr. 1997. An oxygen isotope model for the composition of Mars. *Icarus* 126:373–394.

Lodders K. and Fegley B., Jr. 1998. *The planetary scientist's companion.* New York: Oxford University Press. 371 p.

Longhi J., Knittle E., Holloway J. R., and Wänke H. 1992. The bulk composition, mineralogy, and internal structure of Mars. In *Mars,* edited by Kieffer H. H., Jakosky B. M., Snyder C. W., and Matthews M. S. Tucson: University of Arizona Press. pp. 184–208.

Mao K. K. and Hemley R. J. 1998. New windows on the Earth's deep interior. In *Reviews in mineralogy, vol. 37, Ultrahigh pressure mineralogy: Physics and chemistry of the Earth's deep interior,* edited by Hemley R. J. Washington D.C.: Mineralogical Society of America. pp. 1–32.

Mason B. and Wiik H. B. 1962a. The Renazzo meteorite. *American Museum Novitates* 2106:1–11.

Mason B. and Wiik H. B. 1962b. Descriptions of two meteorites: Karoonda and Erakot. *American Museum Novitates* 2115:1–10.

Mattey D., Lowry D., and Macpherson C. 1994. Oxygen isotope composition of mantle peridotite. *Earth and Planetary Science Letters* 128:231–241.

McDonough W. F. 2001. The composition of the Earth. In *Earthquake thermodynamics and phase transformations in the Earth's interior,* edited by Teisseyre R. and Majewski E. San Diego: Academic Press. pp. 3–23.

McDonough W. F. and Rudnick R. L. 1998. Mineralogy and composition of the upper mantle. In *Reviews in mineralogy, vol. 37, Ultrahigh pressure mineralogy: Physics and chemistry of the Earth's deep interior,* edited by Hemley R. J. Washington D.C.: Mineralogical Society of America. pp. 139–164.

McDonough W. F. and Sun S. S. 1995. The composition of the Earth. *Chemical Geology* 120:223–253.

Meyer C. 2003. *Mars meteorite compendium—2003.* Johnson Space Center, JSC#27672 Rev. B. http://curator.jsc.nasa.gov/antmet/mmc/mmc.htm.

Mohapatra R. K. and Murty S. V. S. 2003. Precursors of Mars: Constraints from nitrogen and oxygen isotopic compositions of martian meteorites. *Meteoritics & Planetary Science* 38:225–241.

Newton J., Franchi I. A., and Pillinger C. T. 2000. The oxygen-isotopic record in enstatite meteorites. *Meteoritics & Planetary Science* 35:689–698.

O'Neill H. St. C. 1991a. The origin of the Moon and the early history of the Earth—A chemical model. Part 1: The Moon. *Geochimica et Cosmochimica Acta* 55:1135–1157.

O'Neill H. St. C. 1991b. The origin of the Moon and the early history of the Earth—A chemical model. Part 2: The Earth. *Geochimica et Cosmochimica Acta* 55:1159–1172.

O'Neill H. St. C. and Palme H. 1998. Composition of the silicate Earth: Implications for accretion and core formation. In *The Earth's mantle: Composition, structure, and evolution,* edited by Jackson I. Cambridge: Cambridge University Press. pp. 3–126.

Poirier J. P. 1994. Light elements in the Earth's outer core: A critical review. *Physics of the Earth and Planetary Interiors* 85:319–337.

Prewitt C. T. and Downs R. T. 1998. High pressure crystal chemistry. In *Reviews in mineralogy, vol. 37, Ultrahigh pressure mineralogy: Physics and chemistry of the Earth's deep interior,* edited by Hemley R. J. Washington D.C.: Mineralogical Society of America. pp. 283–317.

Righter K. 2004. Water in terrestrial planets: Always an oxidant? (abstract #1674). 35th Lunar and Planetary Science Conference. CD-ROM.

Ringwood A. E. 1979. *Origin of the Earth and Moon.* New York: Springer-Verlag. 295 p.

Robinson M. S. and Taylor G. J. 2001. Ferrous oxide in Mercury's crust and mantle. *Meteoritics & Planetary Science* 36:841–847.

Rubin A. E., Warren P. H., Greenwood J. P., Verish R. S., Leshin L. A., Hervig R. L., Clayton R. N., and Mayeda T. K. 2000. Los Angeles: The most differentiated basaltic martian meteorite. *Geology* 28:1011–1014.

Sanloup C., Jambon A., and Gillet P. 1999. A simple chondritic model of Mars. *Physics of the Earth and Planetary Interiors* 112: 43–54.

Schulze H., Bischoff A., Palme H., Spettel B., Dreibus G., and Otto J. 1994. Mineralogy and chemistry of Rumuruti: The first meteorite fall of the new R chondrite group. *Meteoritics* 29:275–286.

Sherman D. M. 1997. The composition of the Earth's core: Constraints on S and Si vs. temperature. *Earth and Planetary Science Letters* 153:149–155.

Solomon S. C., McNutt R. L., Gold R. E., Acuña M. H., Baker D. N., Boynton W. V., Chapman C. R., Cheng A. F., Gloeckler G., Head J. W., III, Krimigis S. M., McClintock W. E., Murchie S. L., Peale S. J., Phillips R. J., Robinson M. S., Slavin J. A., Smith D. E., Strom R. G., Trombka J. I., and Zuber M. T. 2001. The Messenger mission to Mercury: Scientific objectives and implementation. *Planetary and Space Science* 49:1445–1465.

Stevenson D. J. 2003. Mission to Earth's core—A modest proposal. *Nature* 423:239–240.

Stixrude L., Hemley R. J., Fei Y., and Mao H. K. 1992. Thermoelasticity of silicate perovskite and magnesiowüstite and stratification of the Earth's mantle. *Science* 257:1099–1101.

Sun S. S. 1982. Chemical composition and origin of the Earth's primitive mantle. *Geochimica et Cosmochimica Acta* 46:179–192.

Taylor L. A., Nazarov M. A., Shearer C. K., McSween H. Y., Jr., Cahill J., Neal C. R., Ivanova M. A., Barsukova L. D., Lentz R. C., Clayton R. N., and Mayeda T. K. 2002. Martian meteorite Dhofar 019: A new shergottite. *Meteoritics & Planetary Science* 37:1107–1128.

Taylor S. R. and Norman M. D. 1990. Accretion of differentiated planetesimals to the Earth. In *Origin of the Earth,* edited by Newsom H. E. and Jones J. H. New York: Oxford University Press. pp. 29–43.

Treiman A. H. 2003. Chemical compositions of martian basalts (shergottite): Some inferences on basalt formation, mantle metasomatism, and differentiation in Mars. *Meteoritics & Planetary Science* 38:1849–1864.

van Keken P. E., Hauri E. H., and Ballentine C. J. 2002. Mantle mixing: The generation, preservation, and destruction of chemical heterogeneity. *Annual Review of Earth and Planetary Sciences* 30:493–525.

Wänke H. 1981. Constitution of the terrestrial planets. *Philosophical Transactions of the Royal Society of London A* 303:287–302.

Wänke H. and Dreibus G. 1988. Chemical composition and accretion history of terrestrial planets. *Philosophical Transactions of the Royal Society of London A.* 325:545–557.

Weidenschilling S. J. 2000. Formation of planetesimals and accretion of the terrestrial planets. *Space Science Reviews* 92:295–310.

Wetherill G. W. 1994. Provenance of the terrestrial planets. *Geochimica et Cosmochimica Acta* 58:4513–4520.

Wilson L. and Keil K. 1991. Consequences of explosive eruptions on small solar system bodies: The case of the missing basalts on the aubrite parent body. *Earth and Planetary Science Letters* 104: 505–512.

Zipfel J., Scherer P., Spettel B., Dreibus G., and Schultz L. 2000. Petrology and chemistry of the new shergottite Dar al Gani 476. *Meteoritics & Planetary Science* 35:95–106.

APPENDIX

For the CM, CV, H, L, and LL chondrites, average chemical analyses (Jarosewich 1990) for falls, which should be the least weathered samples of these meteorite groups, were used. For the CO and EL chondrites, the average values (Jarosewich 1990) for samples that did not appear significantly weathered were used. For the EH chondrites, only one sample (ALH A77295) was analyzed by Jarosewich (1990), and that data is used here. Analyses of Orgueil (Jarosewich 1990) were used for CI chondrites. Unpublished analyses (Jarosewich, personal communication) for Rumuruti were used for the R chondrites. For CR chondrites, analyses for Renazzo (Mason and Wiik 1962a) were used. Data for CK chondrites were from Karoonda (Mason and Wiik 1962b). Data for CH chondrites were from analyses of NWA 470 (Ivanova et al. Forthcoming).

All data was renormalized so that the sums of all measured elements equaled 100 wt% by dividing by the calculated total after subtracting adsorbed water. Since it was assumed that bound water is due primarily to a weathering product in all meteorites except carbonaceous chondrites, it was subtracted from the total for EH, EL, H, L, LL, and R chondrites before renormalizing the data. Even samples of falls of differentiated meteorites contain bound water (Jarosewich 1990), even though they should contain no pre-terrestrial hydrated minerals. Adsorbed water was measured for all meteorite groups except for the CR (Mason and Wiik 1962a) and CK chondrites (Mason and Wiik 1962b). Average values of Si in the metallic iron were assumed for the EH (3.2 wt%) and the EL (1.3 wt%) chondrites (Keil 1968) and then weighted by the weight percent of metallic iron in these respective meteorites.

The average weight percent of oxygen (Table 4) was calculated from oxide measurements for each meteorite type since each meteorite's contribution to the oxygen isotopic values was weighted by its oxygen concentrations. In wet chemistry, oxygen concentrations are not measured directly but can be inferred by determining which elements are present as oxides and by proportioning oxygen to each of the elements to balance charges. Table 2 compiles the average weight percents of Mg, Al, Si, and Fe for each meteorite type. Weight percents of Mg, Al, and Si, respectively, were converted from the weight percents of MgO, Al_2O_3, and SiO_2. The only exceptions are the weight percents of Si for the EH and EL chondrites where we also include Si that is contained in the metallic iron. The weight percent of Fe was calculated from the weight percents of FeO, Fe_2O_3, Fe (metal), and FeS. The values of Mg/Si, Al/Si, Fe/Si, and Fe/Al are listed for all the meteorite groups.

Table 5 lists the weight percents of FeO, Fe_2O_3, Fe_3O_4, Fe (metal), C, and H_2O that were needed for the redox reactions. The EH and EL chondrites are assumed to contain no FeO. The weight percent of magnetite in CI chondrites was assumed to be 11 wt% (Hyman and Rowe 1983) before adjusting for adsorbed water. Then, the FeO and Fe_2O_3 not in the magnetite was calculated. Only the CI chondrites are assumed to contain Fe_2O_3 and Fe_3O_4. Magnetite is found in other carbonaceous chondrites but usually at concentrations less than 2 wt% (Hyman and Rowe 1982). Also, as can be seen in Equation 2, the assumption that some iron is found as Fe_3O_4 instead of FeO will not noticeably affect the bulk composition of a planet after the redox reactions since very little carbon is used in the reaction to convert Fe_3O_4 to FeO. Only the carbonaceous chondrites are assumed to contain H_2O. Values for SO_3 are not compiled since the reaction that includes SO_3 (Equation 6) never occurred in any of the preliminary runs for the Earth and Mars. Only the CI, CM, and CR chondrites were assumed to contain SO_3 (Jarosewich 1990; Burgess et al. 1991) in these preliminary runs.