Constraints on the structure of the martian interior
determined from the chemical and isotopic systematics of SNC meteorites

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Abstract–The crystallization ages of martian (SNC) meteorites give evidence that martian volcanism has continued until recent times—perhaps until the present. These meteorites also indicate that the mantle source regions of this volcanism are modestly to extremely depleted by terrestrial standards. These 2 observations produce a conundrum. How is it that such depleted source regions have produced basaltic magma for such a long time?

This contribution attempts to quantify the radiogenic heat production in 2 distinct martian mantle source regions: those of the shergottites and nakhlites. Compared to the depleted upper mantle of the Earth (MORB), the nakhlite source region is depleted by about a factor of 2, and the shergottite source region is depleted by a factor of 6. According to current geophysical models, the nakhlite source contains the minimum amount of radioactive heat production to sustain whole-mantle convection and basalt generation over geologic time. A corollary of this conclusion is that the shergottite source contains much too little radioactivity to produce recent (<200 Ma) basalts. A model martian interior with a deep nakhlite mantle that is insulated by a shallow shergottite mantle may allow basalt production from both source regions if the divide between the nakhlite-shergottite mantles acts as a thermal boundary layer. Similarities between lunar and martian isotopic reservoirs indicate that the Moon and Mars may have experienced similar styles of differentiation.

INTRODUCTION

Two events in the last decade have greatly affected how we view Mars, as seen through the lens of the martian (SNC) meteorite collection. The first was the discovery and systematic analysis of the Antarctic basaltic shergottite QUE 94201, which is highly depleted in incompatible trace elements. The second was the recognition that shergottites span a significant range of oxygen fugacity (fO_2) and are often not as "Earthlike" in their redox conditions as had been previously thought. These new findings help explain older observations that were puzzling and raise new questions that require explanation.

In particular, the shergottites appear to have been derived originally from a source region that was highly depleted in heat-producing elements (K, U, and Th). In addition, the shergottites are quite young by meteoritic standards. This raises the question of how such a source region could have continued to produce basalts for so long. Below, I shall attempt to quantify this problem by calculating the heat production rates for the 2 martian mantle source regions that have been sampled (i.e., the nakhlite and shergottite source

regions). The shergottite source region will be shown to have too little radioactivity to drive mantle convection—effectively prohibiting basalt production by adiabatic decompression of hot, rising diapirs. On the other hand, the nakhlite mantle appears to have marginally enough radioactivity to drive a slowly convecting mantle. If the martian mantle is stratified with the nakhlite source region lying beneath an upper, insulating, shergottite source, producing basaltic melts from both may be possible.

QUE 94201 AND ITS SOURCE REGION

Compared to terrestrial basalts, the basaltic shergottite QUE 94201 (hereafter QUE) is highly depleted in extremely incompatible trace elements. This is even more remarkable because, compared to most other SNC meteorites, QUE is rather evolved (Mg# ~38 compared to ~54 for EET 79001A groundmass [QUE, Warren and Kallemeyn 1997; 79001A, Longhi and Pan 1989]). The QUE source region was also highly depleted. The initial $^{87}Sr/^{86}Sr$ and $\varepsilon(^{143}Nd)$ for QUE are 0.7013 and +48, respectively (Borg et al. 1997). This contrasts to a common modern component of the terrestrial

mantle (PREMA), with $^{87}Sr/^{86}Sr$ of 0.7033 and $\varepsilon(^{143}Nd)$ of +9 (Zindler and Hart 1986). Even though the PREMA component is somewhat depleted relative to bulk silicate Earth, the QUE source must be further depleted by a factor of 2 in Rb (relative to Sr) and have a ($[^{147}Sm/^{144}Nd]_{source} \div [^{147}Sm/^{144}Nd])_{chondritic}$) that is a factor of 5 larger than PREMA. This makes the QUE mantle very depleted indeed. This conclusion is strengthened by the observation that bulk Mars is probably slightly enriched in Rb relative to the Earth (Wänke and Dreibus 1988). This makes the lower $^{87}Sr/^{86}Sr$ of the QUE source region even more impressive.

An inference from these observations is that, if the QUE mantle was highly depleted in Rb (and Nd), it was likely also depleted in U, Th, and K (the major radioactive heat-producing elements), since these elements are also highly incompatible. This depletion, coupled with the smaller size of Mars compared to the Earth, makes it difficult to keep the interior of Mars warm. In turn, this makes generation of young basalts difficult. As an endmember case, convective modeling calculations indicate that, in the total absence of internal radioactive heating, Mars would have stopped producing basaltic magma within ~1 Ga of its formation (Kiefer, personal communication).

Many of the martian basalts are young. QUE itself is 330 Ma old (Borg et al. 1997), and QUE is an end member of an array of shergottites, which have ages that range from 575 Ma to ~175 Ma (Nyquist et al. 2002). Therefore, ways must be found to maintain high enough temperatures within the martian mantle to permit volcanism as recently as 150–200 Ma ago. Possibly, the shergottites are not truly representative of martian volcanism, although Hartmann and Neukum (2001) argue, based on crater counts, that individual lava flows in Elysium and Amazonis Planitia are ≤10 Ma. Therefore, we need to find reasonable mechanisms for allowing basaltic volcanism to continue (although possibly episodically) over the entire age of the planet. Again, the difficulty here is the perceived infertility of the martian mantle and the concomitant depletions of those elements that contribute radioactive heat.

Figure 1 shows the above mentioned shergottite array in terms of $^{87}Sr/^{86}Sr$ versus $^{87}Rb/^{86}Sr$, which Jones (1989) and Longhi (1991) have interpreted to be a mixing line. The old "age" of the mixing line implies that the reservoirs that interacted recently have been separated for essentially the age of the solar system (~4.5 Ga). If the trend of Fig. 1 formed because basalts from a depleted martian mantle interacted with an enriched, complementary crust, then all the parent melts of the shergottites might need to originate from a QUE-like source.

Thermally, this could be difficult. The liquidus temperature of QUE (an evolved basalt; Mg# ~38) is ~1170°C (McKay et al. 2002). The liquidus temperature of Eg (groundmass of EET 79001, Lithology A; Mg# ~54) is ~1290°C (Wasylenki et al. 1993). For comparison, the
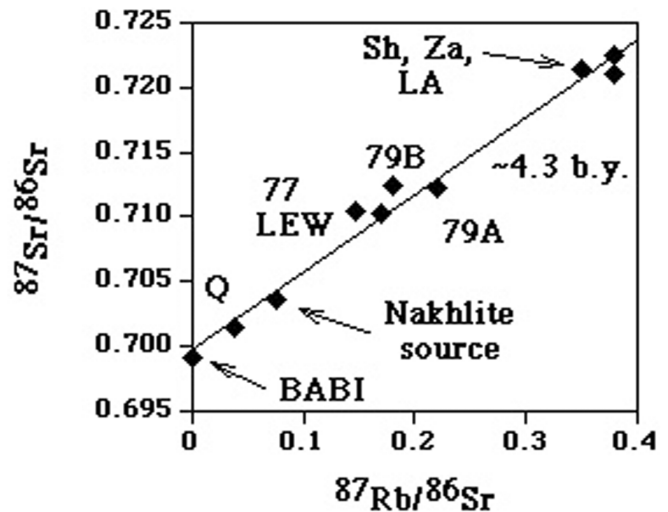

Fig. 1. $^{87}Sr/^{86}Sr$ versus $^{87}Rb/^{86}Sr$. Shergottite Rb-Sr whole rock isochron, after Borg et al. (1997). Initial $^{87}Sr/^{86}Sr$ ratios for the shergottite suite define a ~4.3 Ga "isochron." The source region of the nakhlites and Chassigny at 180 Ma also falls along this trend. Because the shergottites are mainly ~180 Ma in age (with QUE being an exception), this "isochron" is interpreted to be a mixing line between reservoirs that were separated ~4.5 Ga. Within error, the mixing line intercepts the ordinate at the solar system initial value (BABI), again indicating a very early differentiation of Mars. Abbreviations: Sh = Shergotty; Z = Zagami; LA = Los Angeles; 79A and 79B = EET 79001A and EET 79001B; 77 = ALH 77005; LEW = LEW 88516; Q = QUE 94201.

Kilauea tholeiitic lava that formed Makaopuhi Lava Lake, with an Mg# of ~57, had a liquidus temperature of ~1190°C (Wright and Okamura 1977). A primitive MORB composition, V30-RD8-P12, with an Mg# of 61, has a liquidus of ~1235°C (Walker et al. 1979). Thus, despite having lower Mg#s, primitive shergottite magmas tend to have liquidi that are comparable to or higher in temperature than those of terrestrial basalts. Presumably, this difference in liquidus temperatures between the Earth and Mars reflects the more depleted nature of the martian mantle. A depleted mantle could have a higher solidus temperature than a fertile mantle and, consequently, the liquidus temperature of basalts from a depleted mantle might be higher as well.

Thus, the conundrum: How does Mars continue to produce basaltic magmas from source regions that are depleted in radioactive heat sources and that have solidi that are perhaps elevated in proportion to their depletion?

THE NAKHLITE SOURCE REGION

One possible answer to this conundrum is that the QUE mantle is very non-representative and is relatively unimportant to the mass balance of the martian mantle. The example given above of mixing between a single depleted mantle and a crust may be too extreme. Conceivably, many mantle source regions were produced at 4.5 Æ of varying

degrees of depletion. If these can be isolated from each other (to prevent homogenization), then recent melts from such source regions would also produce an array such as Fig. 1. These melts could possibly also interact with crust, just as was envisioned above.

Currently, evidence exists for only a single depleted mantle and 1 enriched crust among the shergottite suite, but this does not consider the nakhlites. Figure 2 (after Borg et al. 2001) shows that the nakhlite source region has had a very different Nd isotopic history from most shergottites. With the exception of Shergotty, the entire shergottite suite is consistent with formation at 4.51 Ga in both the ^{146}Sm-^{142}Nd and ^{147}Sm-^{143}Nd systems (i.e., the 2 systems are concordant if the 2 reservoirs separated at 4.51 Ga). But, this is not true of the nakhlites.

Unlike the shergottites, the nakhlites come from sources that have experienced a more complicated history and fall in a "forbidden" region of Nd isotopic space. This is because the Sm/Nd ratio of the nakhlite source region decreased sometime about 4 Ga ago (Shih et al. 1999). And, consequently, the time-integrated ^{147}Sm/^{144}Nd ratio of the nakhlite source (^{147}Sm/^{144}Nd ~0.23) was considerably smaller than that of the QUE source (^{147}Sm/^{144}Nd ~0.28). This implies that the nakhlite source was less depleted than the QUE source. Possibly, the nakhlite source is more representative of the martian mantle.
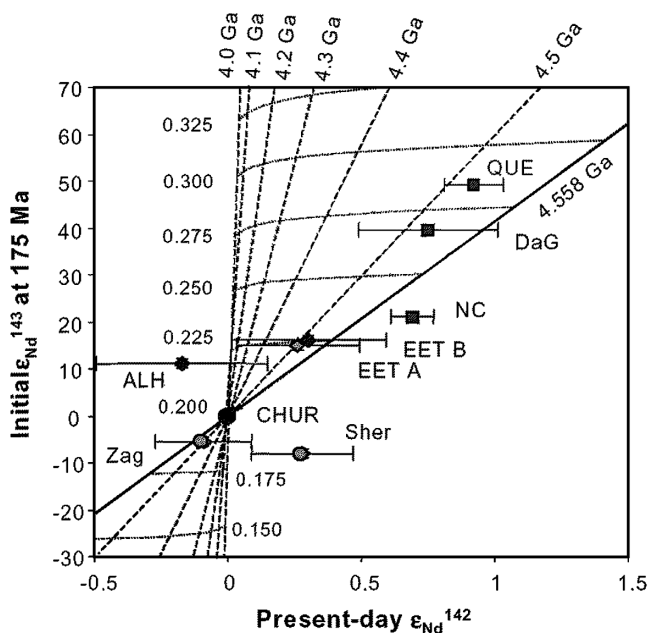

Fig. 2. ε(^{143}Nd) versus ε(^{142}Nd). Condordia diagram for the ^{147}Sm-^{143}Nd and ^{146}Sm-^{142}Nd systems, after Borg et al. (2001). With the sole exception of Shergotty, the shergottites are consistent with derivation from reservoirs that differentiated from chondritic at ~4.51 Ga, in agreement with the Rb-Sr system. The nakhlite-Chassigny (NC) source (and perhaps the Shergotty source) is required to be derived from a reservoir that has undergone a more complex evolution. Abbreviations: Sher = Shergotty; Zag = Zagami; EET A and EET B = EET 79001A and EET 79001B; ALH = ALH 77005; QUE = QUE 94201; DaG = Dar al Gani 476; CHUR = Chondritic values at 175 Ma.

While still depleted, it is not as depleted as the QUE source and may be a more potent heat engine for producing magmas.

I have attempted to estimate the heat productions in the nakhlite and QUE sources and compare them to both primitive chondritic material and the depleted terrestrial upper mantle (MORB). Traditionally, this calculation is performed by analyzing a basalt (e.g., QUE), assuming complete extraction of K, U, and Th into that basalt during melting, and then assuming the degree of partial melting (e.g., 10%). Presumably, though, this method would yield too low abundances of these elements for the QUE source. We know from ^{147}Sm-^{143}Nd systematics that melts we have not sampled were removed from the QUE source before QUE genesis (Borg et al. 1997). Therefore, QUE plus its residuum is not equivalent to the time-integrated QUE source.

Instead, I estimate the composition of the QUE and nakhlite sources using initial ^{87}Sr/^{86}Sr ratios. First, I assume a CaO content for these respective mantles using terrestrial ultramafic rocks as guides. For the QUE source, which I assume to be a harzburgite, I assign a CaO content of 2.2 wt%, and for the nakhlite source, which I assume to be a lherzolite, I assign it 3.0 wt% CaO. Note that the CaO/Sr ratio of the SNC suite is (with some scatter) nearly constant (5.1×10^4) and approaches the chondritic ratio of 6×10^4, implying that Ca and Sr do not strongly fractionate during martian igneous processes (Meyer 2003). The assumed CaO content and the chondritic CaO/Sr ratio are then used to calculate Sr contents for the 2 source regions. Time integrated Rb/Sr ratios for each source (0.013 and 0.027 for QUE and nakhlite sources, respectively) are then used to calculate Rb contents. The relatively constant K/Rb ratio (~200) for the SNC suite is used to calculate K contents. The relatively constant K/U ratio (13,500) is used to calculate the U content. And, finally, the SNC Th/U ratio (~3.5) is used to calculate Th. These abundances are used to calculate watts/gram of heat generation, and these are then compared to CI and MORB sources in Table 1.

The results of this calculation make qualitative sense. The model nakhlite source is about 2.4× as depleted in heat-producing elements as the terrestrial MORB mantle. And in 4.5 Ga, the ε(^{143}Nd) signal generated by the nakhlite mantle is about twice that of MORB. Also, the model QUE mantle is 2.6 times more depleted in heat-producing elements than the nakhlite mantle and has an ε(^{143}Nd) that is ~2.5× larger. Thus, I have generated model mantles with heat-producing capacities that are derived from one isotopic system (Rb/Sr), and these models are qualitatively consistent with another isotopic system (Sm/Nd). Because both of these decay systems are sensitive to the degree of mantle depletion, this consistency is gratifying. Thus, the overall model is internally consistent with isotope geochemistry.

In summary, the conundrum of the previous section may have a solution in the nakhlite source region. Although depleted, the nakhlite source region is several times more radiogenic than the QUE source. This may be enough heat to

Table 1. Calculated heat production in model martian source regions.

Model reservoir	K (ppm)	U (ppb)	Th (ppb)	Heat production (μW/g)[a]	$\varepsilon(^{143}Nd)_{now}$
QUE source	35	2.6	9	6.1×10^{-10}	+50
Nakhlite source	88	6.5	23	1.6×10^{-9}	+20
MORB source[b]	127	22.2	44[c]	3.8×10^{-9}	+10
CI chondrite[b]	558	8.1	29.4	3.5×10^{-9}	$\equiv 0$
Differentiated[d] CI	1140	16.2	60	7×10^{-9}	$\equiv 0$

[a]Computed using the conversion factors of Van Schmus (1995).
[b]Abundances from Newsom (1995).
[c]Assuming MORB Th/U = 2.
[d]Loss of 30 wt% volatiles and a 30 wt% metallic core.

permit recent martian magmatism, especially if the nakhlite source is deep within the planet. Independently, 2D convective modeling indicates that the radioactivity of the model nakhlite source is the minimum amount necessary to sustain whole-mantle convection and basalt production over geologic time (Kiefer 2003).

REDOX VARIATION AMONG THE SHERGOTTITES

Several authors have recently noted that, contrary to popular lore, the shergottites are not completely "Earth-like" in terms of their redox states (Wadhwa 2001; Herd et al. 2001, 2002). Both Eu anomalies in pyroxenes (Wadhwa 2001) and Fe-Ti oxide assemblages (Herd et al. 2001) indicate that the oxygen fugacity (fO_2) of shergottites ranges from near QFM to near or below IW, a range of 3–4 log units.

This observation, coupled with the whole-rock Rb-Sr "isochron" of Fig. 1, indicates recent interaction between 2 reservoirs that have very different redox states and that have been separated for ~4.5 Ga. These 2 reservoirs could be mantle reservoirs, but maintaining good separation over the whole history of the planet is easier if one of the reservoirs is crust or lithosphere. Because Mars apparently lacks subduction-driven plate tectonics (Phillips and Ivins 1979), isolation of crust and mantle is facilitated. Further, fO_2 measurements correlate with geochemical parameters such as $^{87}Sr/^{86}Sr$, $\varepsilon(^{143}Nd)$, and La/Yb (e.g., Fig. 3) that have been associated with crustal contamination (Jones 1989; Longhi 1991; Wadhwa 2001; Herd et al. 2002). The simplest interpretation of these observations is that the shergottite (QUE) mantle is reduced and the crust is oxidized.

As an aside, I note that the discovery of the reduced nature of the martian mantle solves a minor but nagging mystery. Because Mars has a metallic core, we expect that the martian mantle should have been in equilibrium or near equilibrium with iron metal. The FeO contents (~18 wt%) of the SNC meteorites implies that this equilibrium occurred at about IW-1 (Stolper 1977; Walker et al. 1977). Stated differently, because basalts from the Moon, Mars, and the eucrite parent body all have similar FeO contents, we should expect that they have similar redox states of ~IW-1. But, if no subduction of oxidizing materials has occurred over geologic time on Mars, how did the fO_2 of the SNC source regions

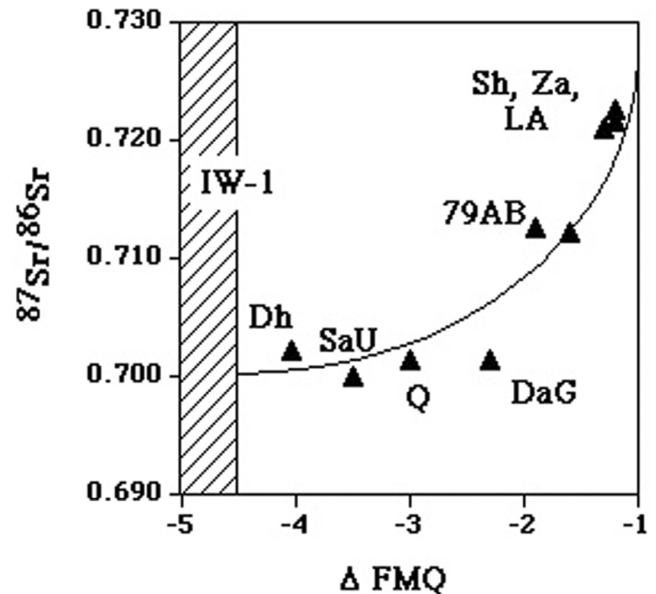

Fig. 3. Correlation between Sr isotopic composition and fO_2. Oxygen fugacity data from Herd et al. (2001, 2002) and Herd (2003). Those shergottites with more primitive $^{87}Sr/^{86}Sr$ ratios are also more reduced. The hatched region is a forbidden zone for rocks of shergottite composition (i.e., FeO content). A simple way to produce the observed trend is for mantle-derived basalts to assimilate old, oxidized crust.

change from IW-1 to QFM? Thus, the discovery that some SNCs formed at or near the anticipated fO_2 of IW-1 makes overall petrologic sense. It also implies that the redox change of some SNCs to near QFM probably occurred after their parent magmas left their source regions. Again, this seems most consistent with the assimilation of oxidized crustal materials. However, the details of this process remain murky (Herd et al. 2002).

COMPARATIVE PLANETOLOGY OF THE SNC SUITE

The above discussion indicates that the QUE source region was reduced to about the same degree as the Moon and the eucrite parent body. Surprisingly, other similarities exist between the Moon and Mars.

Others have emphasized previously that the range of

isotopic variation within the SNC suite is large compared to terrestrial basalts (e.g., Jones 1989). Possibly, some of this difference between the Earth and Mars might be due to differences in bulk composition. For example, Mars is likely to be enriched in alkalis relative to the Earth (e.g., Wänke and Dreibus 1988). A larger bulk Rb/Sr ratio could, in principle, result in larger variation in $^{87}Sr/^{86}Sr$.

Two chronometer systems that ought to be similar for all the terrestrial planets are ^{147}Sm-^{143}Nd and ^{176}Lu-^{176}Hf. These systems are comprised solely of refractory lithophile elements with abundances that should be similar from planet to planet and with bulk planet ratios that should approach chondritic. Therefore, differences or similarities in these chronometer systems between planets should reflect real differences or similarities in the timing and styles of planetary differentiation.

Based on this reasoning, the Moon and Mars appear surprisingly similar. Figure 4 compares terrestrial, lunar, and martian reservoirs in terms of their $\varepsilon(^{143}Nd)$ and $\varepsilon(^{176}Hf)$ signatures. For modern terrestrial basalts, whole rock values are plotted. For the SNCs, their initial ε values are plotted, and the martian crust is calculated by estimating the intersection of the shergottite array (on an isochron diagram) with the 4.5 Æ geochron (e.g., Longhi 1991). For lunar basalts, initial isotopic ratios are used to calculate the parent/daughter ratio of the source region, and this source is then allowed to continue to evolve until the present. In this way, basalts and basalt source regions of similar ages are compared. The isotopic composition of KREEP was calculated using solar system initial ratios and the KREEP composition of Warren and Wasson (1979). This same method was used to calculate the isotopic composition of the average lunar crust of Palme et al. (1991). Terrestrial data are mainly from the Lyon group (Blichert-Toft and Albarede 1997). SNC Hf isotopic data are from Blichert-Toft et al. (1999), and SNC Nd isotopic data are from Borg et al. (1997) and Jones (1989). Lunar Nd and Hf isotopic data are from Unruh et al. (1984) and Beard et al. (1998).

First, Fig. 4 shows that the amount of isotopic diversity in the Moon and Mars is large compared to the Earth. This observation appears to imply that the Moon and Mars differentiated early and did not homogenize afterward. Second, several intriguing similarities exist between lunar and martian samples (Jones 2000; Beard 2002). The QUE, EET 79001, and ALH 77005 sources are surprisingly similar to those of high-Ti mare basalts, and, in addition, the martian "crust" is much like lunar KREEP in terms of its Hf and Nd isotopic compositions. Shergotty and Zagami, which are similar in composition to the soils measured at the Viking and Pathfinder landing sites (McSween 1994), are also similar to the Palme et al. (1991) estimate of bulk lunar crust in Fig. 3. The martian soils have a significant windblown dust component, so they may also approximate an average crustal composition. The only lunar reservoir that is not yet
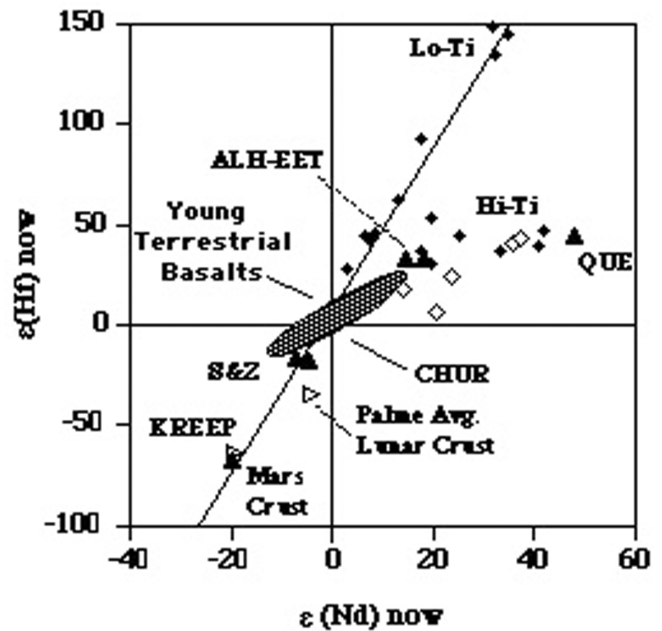

Fig. 4. $\varepsilon(^{176}Hf)$ versus $\varepsilon(^{143}Nd)$ for lunar and martian rocks and reservoirs. The open and filled diamonds are lunar basalt reservoirs calculated for t = 0 (i.e., now); the filled triangles are initial isotopic ratios of martian basalts and crust (Blichert-Toft et al. 1999; Borg et al. 1997; Jones 1989); the open triangles are calculated isotopic ratios for lunar reservoirs from bulk chemical composition, assuming 4.5 Æ of isotopic growth; the shaded ellipse is the field of young terrestrial basalts (Blichert-Toft and Albarede 1997). CHUR is the present-day chondritic ratio of Hf and Nd. Lo-Ti and Hi-Ti refer to the source regions of low-Ti and high-Ti lunar basalts, respectively (Unruh et al. 1984; Beard et al. 1998). The martian basalts plot in the same general region of Nd-Hf isotopic space as the lunar high-Ti basalts and the average lunar crust. The inferred isotopic composition of the martian crust has an isotopic signature that is nearly identical to lunar KREEP. Whatever process produced such reservoirs on the Moon appears to also have acted on Mars.

represented in the martian suite is the low-Ti mare basalt source, which has elevated $\varepsilon(^{176}Hf)$.

A temptation exists to equate the similarities between the Hf and Nd isotopic variations in the Moon and Mars as reflecting similarities in the processes that acted to produce them. Thus, a significant amount of the effort that has been put into understanding lunar differentiation may also be applied to Mars. In particular, if the Moon passed through a magma ocean stage, Mars likely did as well.

A MODEL MARTIAN INTERIOR

Figure 5 gives a schematic example of how the interior of Mars may be structured. First, as a generality, I operate on the principle that various distinct mantle and crustal reservoirs are required to remain isolated from each other over the age of the solar system. Therefore, for this reason, I prefer to interpret the KREEP-like martian component as crust. In a "plum pudding" type of model, the endmember components of Fig. 1 could

possibly be remixed. And, because the fO_2 of primitive shergottites approximately agrees with that expected from metal-silicate equilibrium, I also prefer that the oxidized component of the shergottites be crustal. A physical separation of the shergottite (QUE) mantle and crust prevents isotopic and redox equilibration, which Figs. 1 and 2 seem to require. Therefore, I assume that the enriched, oxidized component of the shergottites is crust and adopt the ~50 km average crustal thickness proposed by Norman (1999) and Zuber et al. (2000).

Other components that need to be isolated over geologic time are the QUE and nakhlite mantles. Time integrated, the QUE source must have been 2–3 times as depleted as the nakhlite source (Table 1). This depletion in LREE and LIL elements also implies depletion in incompatible major elements such as Al, which is necessary to stabilize garnet and spinel. Consequently, because of its depleted nature, the QUE mantle is probably buoyant with respect to the nakhlite mantle. For this reason, I place the shergottite mantle in the upper mantle of Mars. Conceivably, the shergottite mantle is actually lithosphere, although I would prefer that it is convective. This preference arises because a static lithosphere that is periodically metasomatized from below could, in principle, develop a wide range of isotopic components over geologic time. But, Fig. 1 shows evidence of only 2
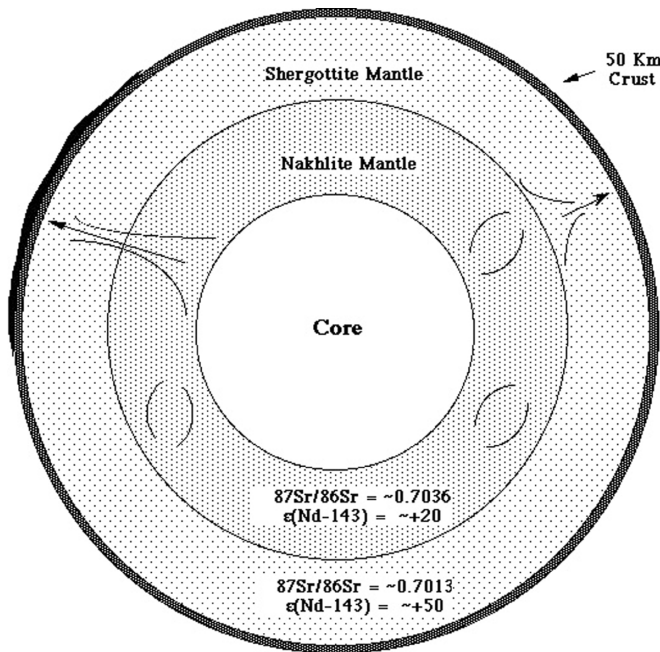

Fig. 5. Schematic of a possible martian interior. The more fertile nakhlite mantle is believed to lie below the shergottite mantle. Because the shergottite mantle is more depleted, it should be more buoyant. The shergottite mantle acts as an insulator to the nakhlite mantle, allowing its heat to be better retained and allowing volcanism to continue until recently. The core-mantle boundary and the shergottite-nakhlite mantle boundary are thought to act as thermal boundary layers. The depths of the shergottite-nakhlite mantle transition and the core-mantle boundary are unknown, and the locations shown in the figure are purely schematic.

components, one of which I have attributed to crust. Therefore, I would prefer that the shergottite mantle be convectively well-mixed.

Placing the nakhlite mantle below the shergottite mantle has the additional benefit of putting the most fertile mantle, with its heat-producing elements, deep within the planet. I anticipate that heat generated at depth could result in plumes from both the core mantle boundary and from the shergottite-nakhlite mantle boundary. Presumably, both these interfaces will act as thermal boundary layers where heat can be exchanged without exchange of mass.

Conceptually, this situation is not very different from the way the Earth's mantle is perceived to operate, except that no provision is made for subducting slabs. Terrestrial oceanic island basalts are often thought to originate in the lower mantle—perhaps at the core-mantle boundary. Mid-ocean ridge basalts are thought to originate in the more depleted upper mantle.

CONCLUSIONS AND DISCUSSION

Working from the SNC meteorite database, a problem clearly exists with generating young basalts from source regions that are so depleted in heat-producing isotopes. The burial of a sufficiently enriched nakhlite source that is insulated by the more depleted shergottite mantle (and a highly enriched crust) may alleviate this problem. Calculations to date indicate that this is the case.

An alternative solution is that of the "plum pudding" model, where enriched pods of fertile mantle are embedded in a matrix of depleted mantle. I favor the layered mantle because of the difficulty of keeping the various isotopic reservoirs separated over solar system history. The best evidence that the shergottite array (Fig. 1) was generated by basalt-crust interaction is that the degree of crust assimilation appears to correlate with redox state—a result that the "plum pudding" model would not have naturally predicted.

One problem with this interpretation of Mars' interior is that we would also anticipate that the nakhlite mantle would be as reduced, or nearly so, as the shergottite mantle. Data on the redox state of the nakhlite mantle conflict. Because of the similarities in Sr and Nd initial isotopic ratios, Jones (1989) assumed that both Chassigny and the nakhlites originated in the same source region. However, nakhlites show little, if any, Eu anomaly, but Chassigny does (Wadhwa 1994; Meyer 2003). The reasons for this discrepancy are not known. Possibly, this means that the schematic of Fig. 5 is oversimplified and that much remains to be learned about Mars and its igneous evolution. This would not be surprising.

Another possible problem with this model is that the shergottite suite has sub-chondritic La/Th ratios (McLennan 2003). I expect that bulk Mars should have a chondritic La/Th ratio and, therefore, should have a complementary reservoir that has La/Th greater than chondritic. In the simplest version

of the model presented here, the complementary reservoir to the depleted shergottite mantle would be the crust. However, if anything, La/Th is lowest for those samples with the greatest amount of the "crustal" component. This low La/Th signature for the crust makes reasonable sense geochemically. Thorium is more incompatible than La, so small degree partial melts or highly fractionated liquids should have smaller La/Th than the bulk system. The real question is why samples such as QUE do not have elevated La/Th. Two possible explanations exist. One is that the complementary reservoir is not the crust but the nakhlite mantle. The nakhlites tend to have super-chondritic La/Th. Because the nakhlites appear to have formed as very small degree partial melts (judging by the change in Sm/Nd from that of the nakhlite source), the relative incompatibilities of La and Th argue that the nakhlite source region may have had even larger La/Th than the nakhlites themselves. Therefore, although no known link presently exists between the shergottite and nakhlite mantles, they are possibly somehow complements of each other (e.g., Longhi 1991).

Another possible explanation is that all shergottites began as high La/Th liquids and that all of them (even QUE) have experienced enough crustal contamination to lower their La/Th below the chondritic value. If that were the case, the problems of producing basalts in the shergottite source region would be even greater than I have suggested above.

Finally, as suggested by McLennan (2003), another martian reservoir may exist that we have yet to sample. This is certainly possible. If this reservoir has a high La/Th ratio, it may also have reduced levels of K, Th, and U (Treiman 1986). Consequently, basalt production in this unsampled reservoir may be more difficult than in the shergottite source region.

At least part of our current lack of understanding must stem from the fact that we are attempting to unravel the history of an entire planet from a limited number of samples. Further, the clustering of cosmic ray exposure ages in the SNC suite (Nyquist et al. 2002) implies that the cratering events that liberated the SNCs from Mars are even more limited. Still, we clearly have no martian samples that are demonstrably from source regions with chondritic abundances of heat-producing elements. Therefore, we need to prepare to understand how depleted source regions have recently produced basalt.

Editorial Handling— Dr. Allan Treiman

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