

Experimental petrology of the basaltic shergottite Yamato-980459: Implications for the thermal structure of the Martian mantle

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Abstract—The Martian meteorite Yamato (Y-) 980459 is an olivine-phyric shergottite. It has a very primitive character and may be a primary melt of the Martian mantle. We have conducted crystallization experiments on a synthetic Y-980459 composition at Martian upper mantle conditions in order to test the primary mantle melt hypothesis. Results of these experiments indicate that the cores of the olivine megacrysts in Y-980459 are in equilibrium with a melt of bulk rock composition, suggesting that these megacrysts are in fact phenocrysts that grew from a magma of the bulk rock composition. Multiple saturation of the melt with olivine and a low-calcium pyroxene occurs at approximately 12 ± 0.5 kbar and 1540 ± 10 °C, suggesting that the meteorite represents a primary melt that separated from its mantle source at a depth of ~100 km. Several lines of evidence suggest that the Y-980459 source underwent extensive melting prior to and/or during the magmatic event that produced the Y-980459 parent magma. When factored into convective models of the Martian interior, the high temperature indicated for the upper Martian mantle and possibly high melt fraction for the Y-980459 magmatic event suggests a significantly higher temperature at the core-mantle boundary than previously estimated.

INTRODUCTION

The Martian meteorite Yamato (Y-) 980459 is a basalt with abundant olivine megacrysts, i.e., an olivine-phyric shergottite. Because Y-980459 is the most magnesian shergottite known, it is the most likely, a priori, to represent a primitive mantle-derived basalt composition. It is therefore crucially important in understanding the origins of Martian basalts and their source mantle.

In this work, we use laboratory experiments to address the question of whether Y-980459 represents a basalt composition or whether it contains extra olivine compared to the true magma. We have conducted a series of crystallization experiments using the Y-980459 bulk composition at Martian upper-mantle pressures to map out the pressure-temperature trajectory of the liquidus and determine the composition of the liquidus phases. The first objective of these experiments was to determine whether Y-980459 is indeed a primary melt of the Martian mantle, which we did by comparing the composition of the first-formed crystalline phases from the experiments with the Y-980459 mineral core compositions and looking for multiple saturation on the liquidus. The second objective, given a positive outcome for the first, was

to characterize the mineralogy, depth, and thermal state of the Martian mantle source of Y-980459. Our results demonstrate that the Y-980459 composition is multiply saturated at 12 kbar, suggesting that the melt separated from its source region at a depth near 100 km. Moreover, the very high melting temperature, which was 1540 °C at the multiple saturation pressure, places significant constraints on the thermal structure of the Martian mantle.

Petrography and Chemistry

The petrography and geochemistry of Y-980459 provide the context for motivating this study and for understanding its results. The petrography and mineral chemistry of Y-980459 suggest that it might have the composition of a primitive magma, essentially undifferentiated since it formed in Mars' mantle. The trace element and isotopic chemistry imply that Y-980459 was generated from a depleted portion of Mars' mantle from which basaltic melt had already been extracted.

Y-980459 consists of large olivine crystals (up to 1.1 mm, as individual euhedra and clusters) and smaller pyroxenes (low and high in Ca, up to 0.4 mm) set in a groundmass (or mesostasis) of basaltic glass, Fe-Ti-Cr oxide

minerals as small euhedra and dendrites, and spherical blobs of quenched Fe-S melt (Greshake et al. 2004). It is composed of 48% pyroxene, 26% olivine, 25% mesostasis, and 1% other minerals (Mikouchi et al. 2003). Unlike the other Martian basalts, Y-980459 contains no plagioclase. These textures suggest that olivine was the first mineral to crystallize, followed by low-Ca pyroxene, high-Ca pyroxene and the euhedral oxides. The mesostasis texture suggests rapid cooling and quenching.

Y-980459 is the most magnesian Martian basalt known, with a bulk Mg# (= molar Mg/(Mg + Fe)) of 0.67 (Koizumi et al. 2004). This bulk Mg# is similar to those of the olivine phyric shergottites Dar al Gani (DaG) 476/479 and Sayh al Uhaymir (SaU) 005/094, which may have such high Mg#s because they contain cumulus olivine (Goodrich 2003). Only the lherzolitic shergottites, which do contain abundant cumulate olivine, have higher bulk Mg#s. Of comparable or greater significance are the olivine cores in Y-980459, which are the most magnesian of all Martian meteorites with Mg#s as high as 0.86 (Greshake et al. 2004). Calculations using the MELTS model of Giorso and Sack (1995) and regressions based on olivine partition coefficients show that liquidus olivine compositions should closely resemble the magnesium composition of the olivine cores (McKay et al. 2004). This concurrence of calculated liquidus-olivine composition with that measured in the natural Y-980459 olivine cores suggests that the olivine phenocrysts are not cumulate; furthermore, it indicates that the bulk composition of Y-980459 may represent that of its parent melt. While the calculated olivine compositions are suggestive of the parental nature of Y-980459, they are dependent on the choice of relevant *D* values and so require experimental confirmation. Below, we show experimentally that the core olivines in Y-980459 were in Fe/Mg chemical equilibrium with the bulk rock composition.

Geochemically, Y-980459 is strongly depleted in incompatible elements like the light rare earth elements (LREE); the Y-980459 bulk has a CI-normalized La/Yb ratio of 0.076 (Dreibus et al. 2003; Shirai and Ebihara 2004). Y-980459 is also depleted in other highly incompatible elements like U, Th, and K (Dreibus et al. 2003). For Y-980459, this evidence for a depleted source is generally consistent with the high Mg# of the bulk and the early olivines in Y-980459. The other depleted shergottites include Queen Alexandra Range 94201, DaG 476, and SaU 005 (e.g., Borg et al. 1997, 2003; Herd et al. 2001; Herd 2003). The depleted shergottites were also typically reduced during crystallization, with oxygen fugacities close to the iron-wüstite buffer (Herd et al. 2001; Wadhwa 2001; Goodrich et al. 2003; McCanta et al. 2004). Y-980459 had a similarly low oxygen fugacity during crystallization (McKay et al. 2004).

Furthermore, this depletion in incompatible elements was a characteristic of the source mantle(s) of the depleted

shergottites, which show extremely high ϵ_{Nd} and low initial $^{87}\text{Sr}/^{86}\text{Sr}$ (e.g., Borg et al. 1997, 2003). These isotope initials require sources that were extremely depleted over a very long time. Y-980459 clearly belongs with this group, as it has a very high $\epsilon_{\text{Nd}} = +44$ and a low $^{87}\text{Sr}/^{86}\text{Sr}_i = 0.70148$ (Shih et al. 2005).

In contrast, other shergottites show variable degrees of enrichment in incompatible elements, even though they derived from sources that were time-averaged depleted. The type example is Shergotty, which lacks magnesian olivine but is slightly enriched in LREE (e.g., Longhi et al. 1992). However, the olivine-phyric basalt Northwest Africa 1068/1110 is enriched in LREE (Barrat et al. 2002). These enriched shergottites are relatively oxidized, near ~QFM-1 (Wadhwa 2001; Herd et al. 2001; McCanta et al. 2004; Herd 2006). The geochemical reservoirs or processes responsible for the enriched shergottites are controversial. It could be an assimilated crustal component (e.g., Jones 2003), a separate mantle reservoir (e.g., Borg and Draper 2003), or a result of mantle metasomatism (Treiman 2003).

EXPERIMENTAL METHODS

Experiments were conducted using a Quickpress non-end-loaded piston cylinder apparatus at the NASA Johnson Space Center. A discussion of general operating procedures for the piston cylinder is given by Dunn (1993). A half-inch diameter pressure assembly comprised of a BaCO₃ cell, MgO internal parts, and a graphite sample capsule was employed. Temperature was monitored with a W5Re/W25Re (C-type) thermocouple. Pressures ranged from 4 to 17 kbar and temperatures from 1410 to 1615 °C. Pressure was calibrated using the melting point of diopside (Boyd and England 1963; Williams and Kennedy 1969). The resulting friction correction to the pressure is 0.5 ± 0.1 kbar below gauge in the pressure range 7 to 15 kbar, consistent with the pressure calibration study of McDade et al. (2002). A temperature profile for the furnace assembly was determined using the spinel growth rate method of Watson et al. (2002). There is a gradient of 10 ± 2 °C from the tip of the thermocouple to the inner top of the charge (where crystal growth was initiated in all run products). The run temperatures quoted here have been corrected for this gradient. Details of the pressure and temperature calibration are given in the Appendix. Temperature and pressure ramps were done "hot piston out," that is, the assembly was first brought up to 10% above the run pressure and then ramped up to run temperature over 30 min. Once the run temperature was reached, the pressure was brought down to run pressure. This technique was used as it minimizes the friction between the cylinder wall and the assembly. More importantly, the friction correction is reproducible with the calibration runs that were ramped to pressure and temperature in the same manner. At the end of the run, the assembly was quenched isobarically. Run

Table 1. Experimental starting material. The second and third columns are published compositions of Yamato-980459 from Greshake et al. (2004) and Misawa (2003), respectively. The target composition is an average of these analyses. The final column is the average composition of experimental glasses from runs that produced only glass.

Oxide	Y-980459	Y-980459	Target composition	Composition as weighed	Experimental glass composition
SiO ₂	49.66	48.63	49.15	49.24	49.47
TiO ₂	0.48	0.54	0.51	0.51	0.50
Al ₂ O ₃	6.03	5.26	5.65	5.67	5.90
Cr ₂ O ₃	0.71	0.71	0.71	0.71	0.68
FeO	15.88	17.29	16.59	16.63	16.00
NiO	0.03	0.03	0.03	0.03	0.01
MnO	0.43	0.52	0.48	0.48	0.55
MgO	18.20	19.61	18.90	18.95	18.91
CaO	7.24	6.36	6.80	6.81	6.85
Na ₂ O	0.80	0.48	0.64	0.64	0.75
K ₂ O	0.02	0.02	0.02	0.02	0.02
P ₂ O ₅	0.31	0.29	0.30	0.30	0.34
S					0.01
FeS	0.19	0.26	0.23	0.23	
Total	100.00	100.00	100.00	100.00	100.00

conditions are shown in Table 2. In addition, a 1-atmosphere melting experiment was performed in a gas-mixing furnace with fO_2 controlled at $IW + 0.7 \pm 0.1$ log units.

The experimental starting composition (Table 1) is an average of the reported whole-rock values (Misawa 2003; Greshake et al. 2004). To make this composition, oxides and carbonates were ground together under acetone with an agate mortar and pestle, then melted completely in a muffle furnace and reground to ensure homogeneity. The starting materials were stored in an oven above 110 °C to keep them dry. In addition to this precaution, experimental assemblies were placed in the same oven for at least an hour prior to being placed into the piston cylinder apparatus. Oxygen fugacity was controlled according to the equilibria described by Holloway et al. (1992): the starting material with a preset Fe₂O₃ content (Kilinc et al. 1983; Kress and Carmichael 1988) is loaded into a graphite capsule. The graphite influences the fO_2 by reacting with the Fe₂O₃ in the melt to form dissolved CO₂ and FeO. The Fe₂O₃ content of the starting material was preset by placing it in a CO/CO₂ gas mixing furnace at 800 °C with log (fO_2) maintained at -14.9 ± 0.1 for 90 h. The resulting Fe₂O₃ concentration in the starting material is calculated to be 1.0 ± 0.1 wt% (corresponding to an Fe³⁺/Fe²⁺ ratio of 0.06), yielding an fO_2 for the piston cylinder runs of $IW + 0.7 \pm 0.2$ log units.

Run products were analyzed using the Cameca SX-100 electron microprobe at NASA/JSC. A one-micron focused electron beam with a current of 15 na and an accelerating voltage of 20 kv was employed for calibrations and analyses. X-rays were analyzed using WDS crystal detectors. Elemental concentrations were calibrated against well-characterized natural mineral and pure metal standards: kaersutite for Si, Fe, Ti, Al, Mg, and Ca; pure metals for Cr and Ni; spessartine garnet for Mn; oligoclase for Na;

orthoclase for K; apatite for P; and troilite for S. Peaks and backgrounds were counted for 20 s for major elements and 60 s for minor elements. A $\phi\rho z$ correction was applied to the data.

RESULTS

Phase assemblages in the experimental run products include glass only, olivine + glass, pyroxene + glass, and olivine + pyroxene + glass (Table 2). A typical run product is shown in Fig. 1; phase compositions are given in Table 3.

Were the Experiments at Equilibrium?

The first task was to determine whether the olivine and melt (glass) in these experiments represent chemical equilibrium, which we tested by comparing the distribution of Fe and Mg between them. Olivine/melt partition coefficients (D_i , or D value, = [atomic concentration element i in olivine]/[atomic concentration element i in basaltic melt]) for Mg and Fe are strongly correlated (Jones 1995; Beattie et al. 1991). In Fig. 2, D_{Mg} and D_{Fe} from our experiments are compared to the equilibrium correlation line of Jones (1995) for equilibrium experiments conducted at pressures ranging from 1 bar to 40 kbar. For a given D_{Mg} , our experiments give a slightly higher D_{Fe} than the nominal correlation of Jones (1995), but well within his 1σ variation uncertainty range. In terms of D_{Fe}/D_{Mg} (Kd_{Fe-Mg}), our experiments average 0.35 versus the 0.32–0.34 of the correlation line. The value of 0.32 is also returned by the algorithm of Toplis (2005) for the Y-980459 bulk and olivine compositions (Tables 1 and 3). The difference between D_{Fe}/D_{Mg} in our experiments and those of Jones (1995) could be ascribed to pressure, as D values from the high-pressure equilibria of Beattie et al. (1991) tend to lie

Table 2. Run conditions and run products. Ol = olivine and pxn = pyroxene. The weight fraction of phases was calculated by applying a least-squares fit to the mass balance of the eight major oxides (SiO₂, TiO₂, Al₂O₃, FeO, MnO, MgO, CaO, and Na₂O) in the glass, starting materials and crystals.

Sample	Pressure (kbar)	Temperature (°C)	Duration (h)	Results	Calculated weight percent			
					Olivine	Pyroxene	Glass	Sum r^2
Y 16-18	17.2 ± 0.7	1615	6	Glass			100	
Y 16-32	15.2 ± 1.0	1580	24	Glass			100	
Y 16-39	14.7 ± 0.45	1580	22	Glass			100	
Y 15-37	13.9 ± 0.4	1565	23	Glass			100	
Y 15-35	13.8 ± 0.4	1530	22	Pxn + glass		31.0	69.6	0.78
Y 14-33	12.6 ± 0.2	1530	24	Pxn + glass		8.8	90.7	0.84
Y 13-38	12.0 ± 0.4	1550	22	Glass			100	
Y 12-17	11.8 ± 0.4	1540	24	Ol + glass	8.5		91.8	0.70
Y 12-31	12.0 ± 0.7	1525	23	Ol + pxn + glass	4.7	3.5	91.3	0.89
Y 12-29	12.0 ± 0.6	1510	20	Ol + pxn + glass	1.8	26.9	70.5	0.45
Y 12-6 ^a	12.2 ± 0.8	1490	6	Ol + pxn + glass	-2.0	31.05	70.2	0.67
Y 8-19	8.2 ± 0.5	1510	18	Glass			100	
Y 8-21	7.5 ± 0.6	1440	22	Ol + pxn + glass	8.1	13.8	77.5	0.89
Y 8-20	7.4 ± 0.5	1480	22	Ol + glass	11.9		88.5	0.71
Y 7-28	7.1 ± 0.4	1410	17	Ol + pxn + glass	4.5	6.5	88.8	0.69
Y 5-24	4.8 ± 0.2	1465	2	Ol + glass	0.8		98.2	0.33
Y 5-2	4.5 ± 1.1	1445	18	Ol + glass	11.1		90.1	0.17
Y 5-25	4.5 ± 0.3	1440	22	Ol + glass	11.5		89.7	0.22
Y 5-27	4.4 ± 0.2	1410	20	Ol + glass	15.2		85.4	0.39
Y 2	0.001	1455	2	Glass			100	

^aGlass and pyroxene compositions are poorly determined for run 12-6 because of the presence of abundant quench pyroxene crystals.

slightly above the correlation line for low- and high-pressure experiments. However, Toplis (2005) found that D_{Fe}/D_{Mg} was only weakly dependent on pressure. At the end, the D_{Mg} and D_{Fe} values for the experimental Y-980459 run products are consistent with the attainment of equilibrium, within uncertainties of the available experimental data sets.

Was Y-980459 a Melt Composition or a Cumulate?

Having established that equilibrium was attained in these experiments, we can test whether Y-980459 represents a bulk-melt composition. If this is the case, the cores of the natural Y-980459 olivines should be in equilibrium with a melt that has the Y-980459 bulk composition. The composition of the olivines from the experimental run products with the lowest crystal fraction should be closest to the composition of the natural olivine cores. This prediction is borne out—the most magnesian olivine produced in any experiment, Fo₈₆ (Table 3), is nearly exactly that of the olivine cores in Y-980459 itself (Koizumi et al. 2004). This result in itself means that Y-980459 is not a cumulate rock, with excess olivine over that of a magma composition. If Y-980459 were a cumulate, the most magnesian olivine in our experiments should be more so than those of Y-980459 itself.

To confirm this observation, one should find that experimental charges with the most magnesian olivine should have the lowest proportions of olivine crystals—otherwise, mass balance is not satisfied. The compositions of olivines are

measured as above. The fraction of crystals in an experimental charge is calculated as a least-squares fit to mass balance on the eight major oxides (SiO₂, MgO, FeO, CaO, Al₂O₃, TiO₂, MnO, and Na₂O) among the starting materials (Table 1) and product glasses and crystals (Table 3). The results of these calculations are given in Table 2 along with the r^2 value (the square of the sum of the residuals, or difference between the calculated and actual bulk-oxide values). All runs give r^2 values below 1.0, indicating good mass balance. These results are graphed in Fig. 3, which shows that Mg# of olivine is inversely correlated with the fraction of crystals present. The olivine with the highest Mg# = 0.86 is in the experimental run with the lowest fraction of crystals (run Y5-24), and is the same as that for the cores of the olivines in Y-980459 (Koizumi et al 2004).

The fact that the magnesium contents for the first-formed experimental olivines and the cores of the natural olivines in Y-980459 are the same indicates that the natural olivine cores were in equilibrium with a melt that was compositionally identical to the Y-980459 bulk composition. This is strong evidence that Y-980459 is not a cumulate, and therefore its bulk composition is representative of the bulk melt composition from which it formed.

At What P-T Did the Y-980459 Magma Form?

The temperature and pressure conditions (P-T) of the crystallization experiments and resulting phase assemblages

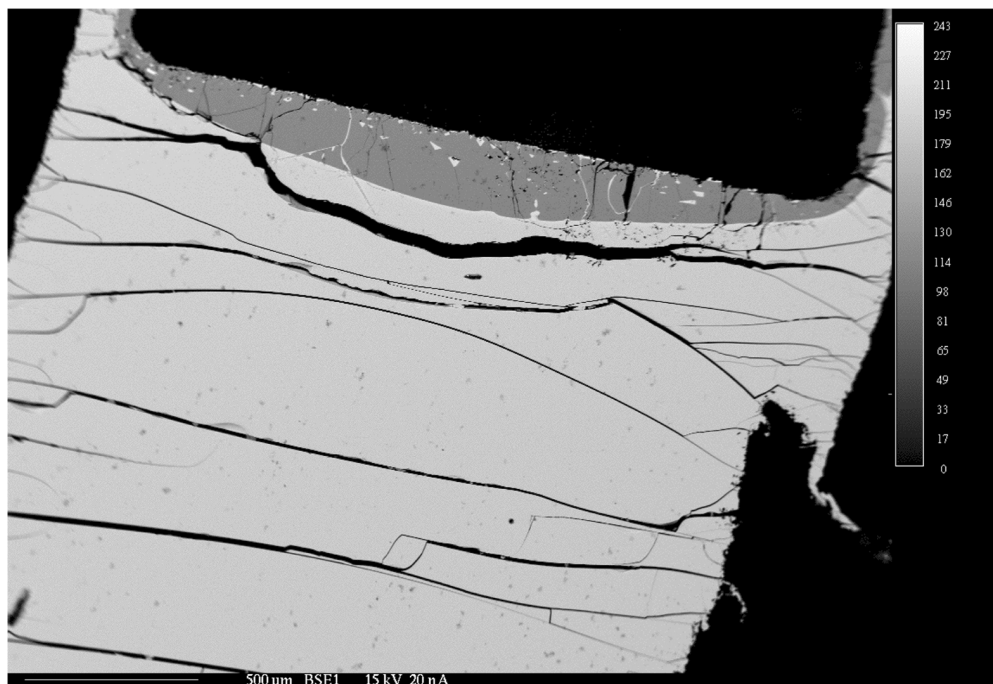


Fig. 1. A typical run product. Backscattered electron image of the sample charge from run Y12-17, run at 1540 °C and 11.8 ± 0.4 kbar. Olivine crystals are clustered below the cap of the graphite capsule.

are plotted in Fig. 4 with the liquidus, olivine-in line, and pyroxene-in line. Olivine is the liquidus phase for Y-980459 from 1 bar up to 12 ± 0.5 kbar pressures and 1540 ± 10 °C, above which point the liquidus phase is low-Ca pyroxene (Table 2; Fig. 4). This is the liquidus co-saturation point for the Y-980459 composition. At that point alone can Y-980459 melt coexist with the mantle mineral assemblage olivine + low-Ca pyroxene, and that point is the most likely (the only) P-T at which the Y-980459 magma could last have equilibrated with a polyminerally mantle. This is consistent with the Y-980459 magma melting at 12 kbar and 1540 °C, or with a deeper magma stopping and equilibrating at that point. Intriguingly, nearly this same pressure appears for the olivine + low-Ca pyroxene co-saturated liquidus of a Gusev basalt composition, 10 kbar at 1310 °C (Monders et al. 2006).

Norris and Herd (2006) presented preliminary experimental phase equilibria data using a synthetic Y-980459 analog. They estimate the multiple saturation point at 15 kbar and 1490 °C, or about 50 °C lower temperature and 3 kbar higher pressure than the results from this study. This is consistent with the fact that their Y-980459 analog has a higher FeO (18 cf 16 wt%) than that used in this study and contained a modest amount of water due to the use of a talc-Pyrex assembly.

DISCUSSION

The features of the phase relations for Y-980459 discussed above have important implications for its

petrogenesis. First, the existence of multiple saturation of the bulk composition is a strong argument for Y-980459 being a primary mantle melt. Melts that have fractionated following separation from the mantle by fractional crystallization will no longer be compositionally in equilibrium with the mineral assemblage of their source region. Second, assuming that the melt was eutectic, the temperature and pressure of multiple saturation give us the conditions of the mantle source region. Assuming an average crustal thickness and density of 50 km and 2900 kg/m³ (Zuber et al. 2000; Neuman et al. 2004) and an upper mantle density of 3400 kg/m³ (Bertka and Fei 1997), the pressure of multiple saturation (12 kbar) indicates separation from the Y-980459 mantle source at a depth of about 100 km. If the Y-980459 melt formed in an ascending mantle plume (Kiefer 2003; Redmond and King 2004), 100 km may represent the thickness of the thermal lithosphere.

The Y-980459 melt was likely produced over a range of pressures in a rising mantle plume that stalled out at the base of the lithosphere. The Y-980459 melt then separated from its mantle source at this depth. The phases at the multiple saturation point are interpreted as those which were in equilibrium with the Y-980459 melt at the time of magma separation. Thus, the phases on the liquidus give us the residual mineralogy of the mantle source, after melting, at the depth of magma separation. Only two phases—olivine and low-Ca pyroxene—are on the liquidus at the point of multiple saturation. The absence of a major Ca- and Al-bearing phase, such as plagioclase or minor oxide phases, on the liquidus

Table 3. Composition of the run products from electron microprobe analysis.

Runs with glass only														
	Y 2		Y 16-18		Y 8-19		Y16-32		Y 15-37		Y 13-38		Y 16-39	
	Glass		Glass		Glass		Glass		Glass		Glass		Glass	
	(28)	±	(12)	±	(12)	±	(11)	±	(6)	±	(60)	±	(7)	±
SiO ₂	50.71	0.32	49.30	0.19	49.59	0.16	50.80	0.48	46.55	0.60	49.10	0.43	48.67	0.16
TiO ₂	0.50	0.02	0.50	0.03	0.52	0.04	0.52	0.03	0.49	0.08	0.50	0.05	0.44	0.05
Al ₂ O ₃	5.40	0.05	5.44	0.05	5.47	0.03	5.52	0.08	8.53	0.18	5.58	0.08	5.19	0.04
Cr ₂ O ₃	0.68	0.06	0.69	0.04	0.70	0.03	0.53	0.04	0.80	0.11	0.68	0.04	0.66	0.06
FeO	16.28	0.36	16.19	0.34	16.29	0.29	16.06	0.19	14.13	0.69	16.42	0.13	16.14	0.08
NiO	0.03	0.02	0.00	0.00	0.00	0.00	0.00	0.00	0.01	0.02	0.00	0.03	0.04	0.04
MnO	0.53	0.06	0.46	0.05	0.47	0.06	0.49	0.04	0.61	0.09	0.67	0.03	0.64	0.03
MgO	18.55	0.44	18.86	0.11	18.98	0.28	17.83	0.11	19.82	0.80	18.64	0.12	19.09	0.08
CaO	6.91	0.15	6.71	0.04	6.75	0.07	6.92	0.08	6.66	0.28	6.85	0.06	6.92	0.05
Na ₂ O	0.78	0.08	0.81	0.01	0.82	0.03	0.67	0.05	0.41	0.10	0.84	0.02	0.87	0.03
K ₂ O	0.02	0.00	0.02	0.00	0.02	0.00	0.02	0.01	0.01	0.02	0.02	0.01	0.02	0.01
P ₂ O ₅	0.36	0.01	0.36	0.00	0.36	0.02	0.35	0.00	0.29	0.06	0.34	0.02	0.34	0.01
SO ₂	0.00	0.00	0.01	0.00	0.01	0.01	0.01	0.00	0.02	0.02	0.02	0.02	0.04	0.01
Total	100.76	0.42	99.36	0.44	99.98	0.45	99.71	0.32	98.32	1.07	99.68	0.43	99.07	0.20
Mg#	0.67		0.68		0.68		0.66		0.71		0.67		0.68	

Runs with pyroxene + glass only								
	Y 14-33				Y 15-35			
	Pyroxene		Glass		Pyroxene		Glass	
	(33)	±	(12)	±	(20)	±	(20)	±
SiO ₂	56.02	0.31	48.75	0.34	54.14	0.94	46.65	1.22
TiO ₂	0.04	0.03	0.52	0.06	0.08	0.03	0.66	0.07
Al ₂ O ₃	0.72	0.05	5.63	0.07	2.11	0.29	6.63	0.20
Cr ₂ O ₃	0.70	0.04	0.69	0.04	1.15	0.12	0.58	0.06
FeO	9.58	0.11	17.18	0.17	11.39	0.19	18.47	0.89
NiO	0.01	0.02	0.01	0.03	0.02	0.03	0.01	0.03
MnO	0.44	0.04	0.69	0.03	0.53	0.05	0.72	0.06
MgO	31.22	0.20	17.69	0.13	28.25	0.43	14.27	0.50
CaO	1.13	0.05	7.16	0.08	1.78	0.08	8.47	0.31
Na ₂ O	0.06	0.01	0.99	0.02	0.09	0.02	1.31	0.07
K ₂ O	0.01	0.01	0.03	0.02	0.00	0.01	0.03	0.01
P ₂ O ₅	0.00	0.01	0.36	0.02	0.01	0.02	0.45	0.02
SO ₂	0.02	0.02	0.02	0.02	0.03	0.01	0.02	0.03
Total	99.97	0.39	99.73	0.46	99.57	1.19	98.28	3.26
Mg#	En ₈₄ Wo ₂		0.65		En ₇₉ Wo ₄		0.58	

Table 3. *Continued.* Composition of the run products from electron microprobe analysis.

Runs with olivine + glass only												
	Y 5-2				Y 12-17				Y 8-20			
	Olivine		Glass		Olivine		Glass		Olivine		Glass	
	(18)	±	(27)	±	(52)	±	(20)	±	(34)	±	(25)	±
SiO ₂	39.01	0.74	49.75	0.74	38.88	0.36	49.93	0.22	39.60	0.30	50.07	0.15
TiO ₂	0.01	0.01	0.58	0.03	0.01	0.01	0.48	0.03	0.01	0.01	0.51	0.02
Al ₂ O ₃	0.03	0.01	6.42	0.05	0.04	0.01	5.75	0.05	0.04	0.01	6.46	0.03
Cr ₂ O ₃	0.48	0.05	0.71	0.06	0.39	0.04	0.65	0.05	0.43	0.04	0.65	0.07
FeO	16.32	0.33	15.85	0.33	14.63	0.29	16.33	0.24	15.57	0.27	16.76	0.20
NiO	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.03	0.02	0.01	0.01
MnO	0.44	0.05	0.51	0.05	0.38	0.04	0.54	0.06	0.38	0.06	0.50	0.05
MgO	43.82	0.28	15.54	0.17	44.68	0.23	16.34	0.12	43.78	0.22	15.32	0.06
CaO	0.22	0.01	7.87	0.06	0.26	0.02	7.87	0.05	0.23	0.02	7.73	0.04
Na ₂ O	0.01	0.01	1.01	0.03	0.02	0.01	1.07	0.03	0.02	0.01	1.10	0.04
K ₂ O	0.00	0.00	0.03	0.00	0.00	0.00	0.02	0.00	0.00	0.00	0.03	0.00
P ₂ O ₅	0.05	0.03	0.42	0.01	0.01	0.01	0.39	0.01	0.00	0.01	0.40	0.01
SO ₂	0.00	0.00	0.00	0.00	0.00	0.00	0.03	0.00	0.01	0.00	0.34	0.01
Total	100.42	0.85	98.71	0.61	99.32	0.67	99.43	0.33	100.11	0.37	99.87	0.29
Mg#	Fo ₈₃		0.64		Fo ₈₄		0.64		Fo ₈₃		0.62	
Kd ^a	0.365				0.328				0.325			
Runs with olivine + glass only												
	Y 5-24				Y 5-25				Y 5-27			
	Olivine		Glass		Olivine		Glass		Olivine		Glass	
	(5)	±	(10)	±	(31)	±	(10)	±	(14)	±	(7)	±
SiO ₂	40.03	0.25	49.98	0.23	39.57	0.12	50.20	0.21	39.44	0.17	50.83	0.43
TiO ₂	0.01	0.02	0.49	0.04	0.01	0.02	0.55	0.04	0.01	0.01	0.61	0.03
Al ₂ O ₃	0.03	0.01	6.00	0.08	0.03	0.01	7.31	0.06	0.02	0.01	6.19	0.06
Cr ₂ O ₃	0.40	0.01	0.65	0.01	0.47	0.01	0.65	0.01	0.49	0.01	0.69	0.01
FeO	13.36	0.09	15.68	0.12	15.59	0.18	15.08	0.09	16.48	0.14	15.58	0.15
NiO	0.01	0.01	0.00	0.01	0.01	0.01	0.00	0.01	0.01	0.01	0.00	0.01
MnO	0.34	0.01	0.50	0.02	0.40	0.02	0.49	0.02	0.41	0.01	0.51	0.02
MgO	44.94	0.12	18.73	0.08	43.40	0.22	14.73	0.10	42.72	0.46	14.37	0.05
CaO	0.21	0.01	6.59	0.07	0.22	0.01	7.52	0.05	0.22	0.02	7.96	0.08
Na ₂ O	0.01	0.00	0.78	0.01	0.01	0.01	1.22	0.01	0.02	0.04	0.92	0.04
K ₂ O	0.00	0.00	0.02	0.01	0.00	0.01	0.04	0.01	0.01	0.01	0.03	0.01
P ₂ O ₅	0.03	0.01	0.33	0.01	0.04	0.02	0.37	0.01	0.04	0.03	0.40	0.00
SO ₂	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.00	0.01	0.01	0.01	0.00
Total	99.37	0.25	99.75	0.24	99.77	0.30	98.18	0.22	99.90	0.41	98.11	0.62
Mg#	Fo ₈₆		0.68		Fo ₈₃		0.64		Fo ₈₂		0.62	
Kd ^a	0.355				0.351				0.36			

Table 3. *Continued.* Composition of the run products from electron microprobe analysis.

Runs with olivine + pyroxene + glass												
Y 12-6 ^b							Y 8-21					
	Olivine (5)	±	Pyroxene (20) ^c	±	Glass (6) ^c	±	Olivine (54)	±	Pyroxene (43)	±	Glass (24)	±
SiO ₂	37.45	0.27	54.21	0.67	47.32	0.30	39.21	0.31	55.74	0.62	49.56	0.20
TiO ₂	0.01	0.01	0.09	0.02	0.62	0.04	0.01	0.01	0.06	0.02	0.61	0.03
Al ₂ O ₃	0.10	0.01	2.16	0.62	6.33	0.04	0.03	0.01	0.57	0.07	6.49	0.05
Cr ₂ O ₃	0.38	0.06	1.16	0.22	0.58	0.02	0.47	0.05	0.76	0.06	0.69	0.06
FeO	19.65	0.13	11.52	0.15	18.40	0.06	17.80	0.35	11.02	0.33	17.28	0.27
NiO	0.68	0.08	0.13	0.12	0.03	0.04	0.05	0.02	0.02	0.02	0.01	0.02
MnO	0.46	0.01	0.40	0.02	0.53	0.01	0.41	0.05	0.40	0.05	0.50	0.05
MgO	39.85	0.23	28.32	0.40	15.22	0.13	41.52	0.57	29.53	0.50	14.38	0.08
CaO	0.31	0.01	2.07	0.27	8.86	0.05	0.24	0.02	1.48	0.08	8.23	0.04
Na ₂ O	0.04	0.01	0.08	0.02	1.25	0.02	0.01	0.01	0.06	0.07	1.09	0.07
K ₂ O	0.01	0.01	0.00	0.00	0.03	0.00	0.00	0.00	0.01	0.01	0.03	0.00
P ₂ O ₅	0.04	0.01	0.01	0.01	0.48	0.01	0.03	0.03	0.00	0.00	0.42	0.01
SO ₂	0.01	0.01	0.01	0.01	0.00	0.01	0.01	0.01	0.00	0.00	0.00	0.00
Total	98.99	0.21	100.16	1.01	99.66	0.42	99.81	0.71	99.65	1.12	99.30	0.31
Mg#	Fo ₇₈		En ₇₈ Wo ₄		0.60		Fo ₈₁		En ₈₀ Wo ₃		0.60	
Kd	0.408						0.357					
Runs with olivine + pyroxene + glass												
Y 12-29							Y 12-31					
	Olivine (20)	±	Pyroxene (40)	±	Glass (10)	±	Olivine (8)	±	Pyroxene (8)	±	Glass (11)	±
SiO ₂	38.72	0.53	55.18	0.52	47.90	0.51	39.86	0.30	56.78	0.66	49.73	0.39
TiO ₂	0.00	0.03	0.05	0.03	0.71	0.05	0.01	0.03	0.03	0.03	0.52	0.04
Al ₂ O ₃	0.08	0.01	1.21	0.13	7.11	0.09	0.04	0.01	0.58	0.01	5.55	0.07
Cr ₂ O ₃	0.41	0.04	0.86	0.07	0.58	0.05	0.38	0.02	0.61	0.03	0.66	0.04
FeO	19.34	0.11	11.37	0.25	18.17	0.16	14.79	0.19	9.55	0.07	16.78	0.10
NiO	0.38	0.07	0.06	0.03	0.00	0.04	0.08	0.03	0.04	0.01	0.01	0.03
MnO	0.42	0.04	0.37	0.03	0.51	0.01	0.34	0.02	0.35	0.01	0.50	0.03
MgO	40.88	0.38	29.53	0.34	14.14	0.16	43.90	0.19	30.26	0.23	17.06	0.16
CaO	0.29	0.02	1.60	0.09	8.79	0.09	0.22	0.01	1.34	0.03	7.18	0.05
Na ₂ O	0.02	0.01	0.05	0.01	0.98	0.06	0.01	0.00	0.07	0.02	0.68	0.02
K ₂ O	0.00	0.01	0.00	0.01	0.03	0.01	0.00	0.00	0.00	0.01	0.02	0.01
P ₂ O ₅	0.08	0.06	0.01	0.01	0.49	0.02	0.04	0.03	0.01	0.01	0.37	0.02
SO ₂	0.00	0.02	0.00	0.02	0.00	0.01	0.01	0.01	0.00	0.01	0.01	0.01
Total	100.64	0.50	100.30	0.22	99.44	0.52	99.70	0.41	99.61	0.87	99.08	0.34
Mg#	Fo ₇₉		En ₈₀ Wo ₃		0.58		Fo ₈₄		En ₈₀ Wo ₃		0.64	
Kd	0.368						0.342					

Table 3. *Continued.* Composition of the run products from electron microprobe analysis.

Runs with olivine + pyroxene + glass						
Y 7-28						
	Olivine (5)	±	Pyroxene (22)	±	Glass (7)	±
SiO ₂	39.61	0.09	55.17	0.43	49.36	0.51
TiO ₂	0.00	0.01	0.06	0.03	0.54	0.03
Al ₂ O ₃	0.04	0.02	0.75	0.29	5.73	0.06
Cr ₂ O ₃	0.44	0.01	0.77	0.13	0.67	0.01
FeO	15.69	0.11	10.65	0.62	16.75	0.15
NiO	0.01	0.01	-0.01	0.02	0.00	0.01
MnO	0.39	0.01	0.39	0.02	0.51	0.02
MgO	43.45	0.20	29.69	0.77	16.61	0.17
CaO	0.21	0.02	1.62	0.23	7.47	0.10
Na ₂ O	0.00	0.01	0.05	0.02	0.84	0.02
K ₂ O	0.00	0.00	0.00	0.01	0.02	0.01
P ₂ O ₅	0.01	0.01	0.01	0.01	0.38	0.01
SO ₂	0.00	0.01	0.01	0.01	0.01	0.01
Total	99.88	0.26	99.18	0.16	98.89	0.31
Mg#	Fo ₈₃		En ₇₉ Wo ₃		0.64	
Kd	0.358					

^aKd values are for D_{Fe}/D_{Mg} (olivine/melt); D_i (concentration of element i in olivine)/(concentration of element i in glass).

^bGlass and pyroxene compositions are poorly determined for run 12-6 because of the presence of abundant quench pyroxene crystals.

^cMg# (for glasses) = molar Mg/(Mg + Fe).

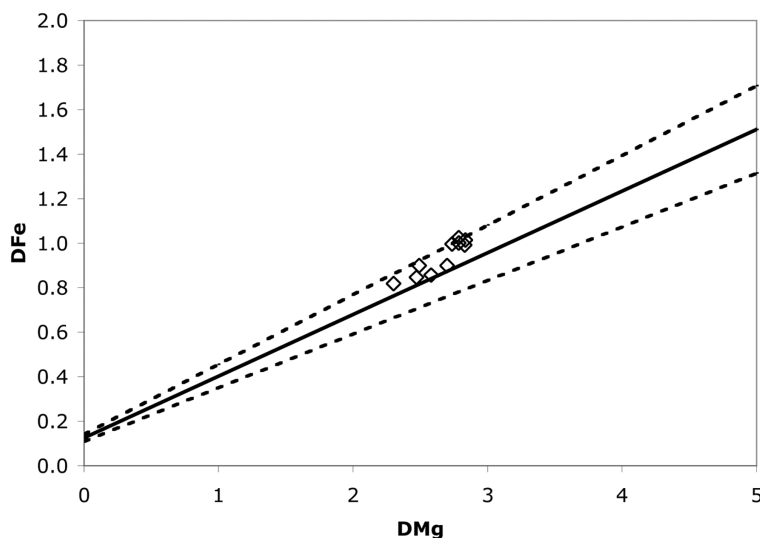


Fig. 2. The plot of molar crystal/melt partition coefficients D_i (mole fraction element i in crystal/mole fraction element i in melt) for Mg and Fe for olivine in the run products of the Y-980459 crystallization experiments. The solid line is the best-fit regression line for equilibrium crystallization experiments (after Jones 1995). The dashed lines are the 1σ variation of the equilibrium experiments from the regression line. Run 12-6 is not included in plot because glass composition is poorly determined due to abundant quench pyroxenes.

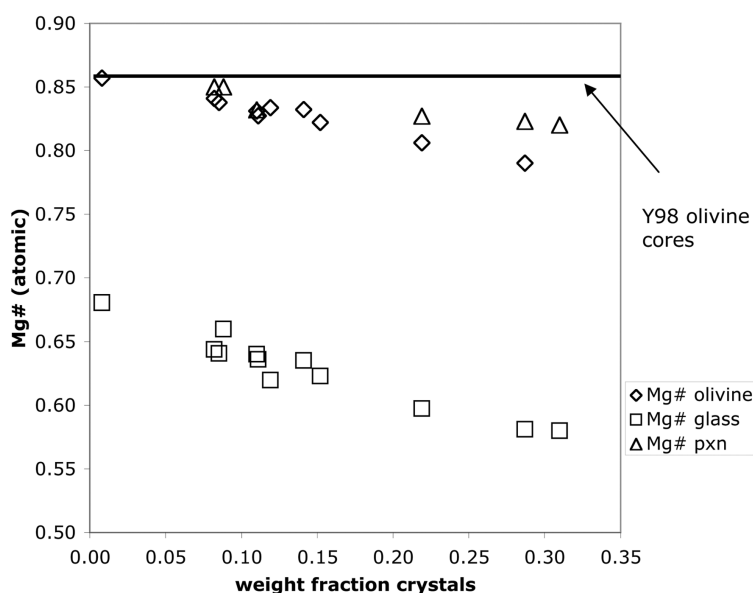


Fig. 3. The plot of Mg# (molar Mg/(Mg + Fe)) versus total weight fraction of crystals for the run products of the Y-980459 crystallization experiments. The solid horizontal line is the composition of the olivine cores in Y-980459. Weight fraction of crystals calculated by least squares mass balance.

suggests a fairly high degree of partial melting, which consumed the lower modal abundance phases in the source region. Shih et al. (2005) noted that almost no Rb/Sr fractionation occurred at the times of depleted shergottite (including Y-980459) formation. They suggested that this was a result of a large degree of partial melting, leaving residual phases that exclude Rb and Sr equally, such as olivine and orthopyroxene, consistent with our results showing only olivine and low-Ca pyroxene on the liquidus at the multiple saturation point. A high degree of partial melting

is also consistent with the high liquidus temperature at the multiple saturation point.

Partial Melting Models for the Martian Mantle

The high magnesium and low silica contents of Y-980459 and other olivine-phyric basalts also suggest a high degree of partial melting and perhaps multiple stages of melt production. By considering the degree of melting needed to achieve the Y-980459 composition, we can place constraints

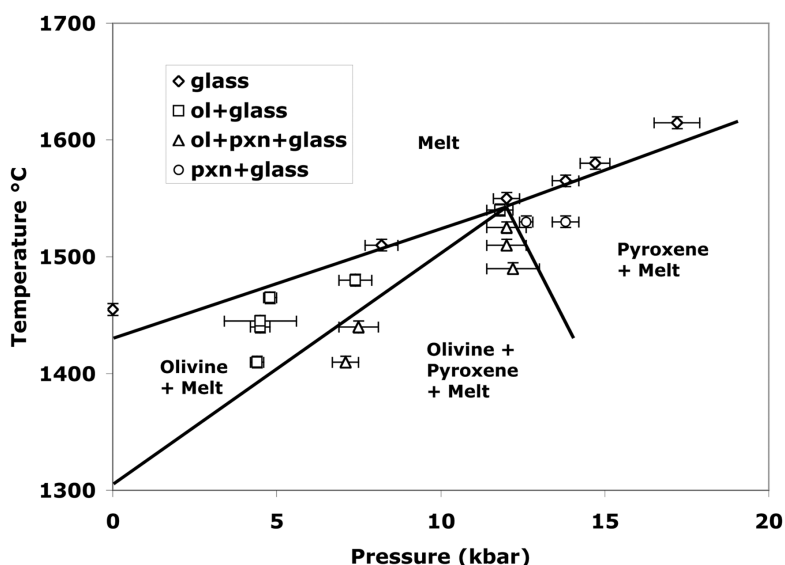


Fig. 4. Experimentally determined phase relations for synthetic Yamato-980459 bulk composition. Diamonds are experiments that produced only glass, squares are experiments that produced olivine + glass, triangles are experiments that produced olivine + pyroxene + glass, and circles are experiments that produced pyroxene + glass. Experimentally determined olivine-in, pyroxene-in and liquidus lines constructed from these results.

on the Y-980459 source mantle as well as on the Mg# of the Martian mantle as a whole. Given that Y-980459 is a primary mantle melt, we can estimate the degree of partial melting necessary to produce the Y-980459 Mg/Fe ratio from mass balance considerations from the bulk Mg# for Y-980459 (0.67), the composition of the first-formed experimental olivines (Mg# = 0.86), and appropriate values for olivine/melt partition coefficients for Fe and Mg (Jones 1995). The composition of the olivines in the Y-980459 source in equilibrium with the Y-980459 melt is the same as the composition of the first-formed experimental olivines. Whatever the Mg# of the protolith (i.e., the Y-980459 source prior to melting), the residual olivines in equilibrium with the Y-980459 melt will have an Mg# = 0.86. For simplicity, we consider only olivine/melt partitioning in estimating the degree of melting of the Y-980459 source. The residuum was likely mostly olivine, so we avoid the complexity of considering the olivine to pyroxene ratio.

Figure 5a shows a plot of the fraction of melt needed to produce the Y-980459 magma as a function of the assumed bulk Mg# for the Y-980459 mantle source prior to production and extraction of the Y-980459 melt, assuming batch melting (Shaw 1970) (see figure caption for model equations and parameter values used). The lower the Mg# of the Y-980459 source prior to melting, the greater the degree of melting required to produce the Y-980459 melt. Consider first the extremes in terms of the possible Y-980459 mantle source composition. A lower limit for the Mg# of the Y-980459 mantle source can be taken to be 0.75, the lowest estimated value for the primitive bulk Martian mantle calculated by Longhi et al. (1992). Assuming this value, a melt fraction of

0.73 is necessary to attain the Mg composition of the residual Y-980459 mantle source. At the other extreme, an infinitesimal melt fraction is possible if the Y-980459 source mantle has the same Mg# (0.86) of the first-formed olivines prior to melting. Intermediate between these extremes of Mg# and melt fraction (see the dashed tie-lines in Fig. 5a), an Mg# = 0.85 for the Y-980459 source requires 10% melting, an Mg# = 0.83 requires 25% melting, an Mg# = 0.81 requires 40% melting, and an Mg# = 0.80 requires 50% melting.

The effects of previous episodes of melting on the Mg/Fe ratio of the Y-980459 mantle source can be illustrated by considering the effect of the fraction of melt extracted on the composition of the residuum (i.e., the effect of a depletion event on the composition of the resulting depleted mantle). A two-stage process—production of magma from a primitive mantle resulting in an increase in the Mg# for the residuum, followed by production and extraction of the Y-980459 magma from this residuum—is assumed in the following example. This simplified two-stage process is used to illustrate the accumulatively large amount of melting necessary to produce the Y-980459 composition and the effect of this melting on the Y-980459 source region. Several episodes of magma production from the mantle source that eventually produced the Y-980459 magma may have occurred prior to the Y-980459 magmatic event. Figure 5b shows a plot of the Mg# of the residual mantle resulting from a given fraction of melt being removed from the Martian mantle (or, conversely, the amount of melt required to be removed from the mantle to achieve a given Mg# of the resulting depleted mantle). As in Fig. 5a, a batch-melting model is assumed. The curve to the upper left is for a Martian

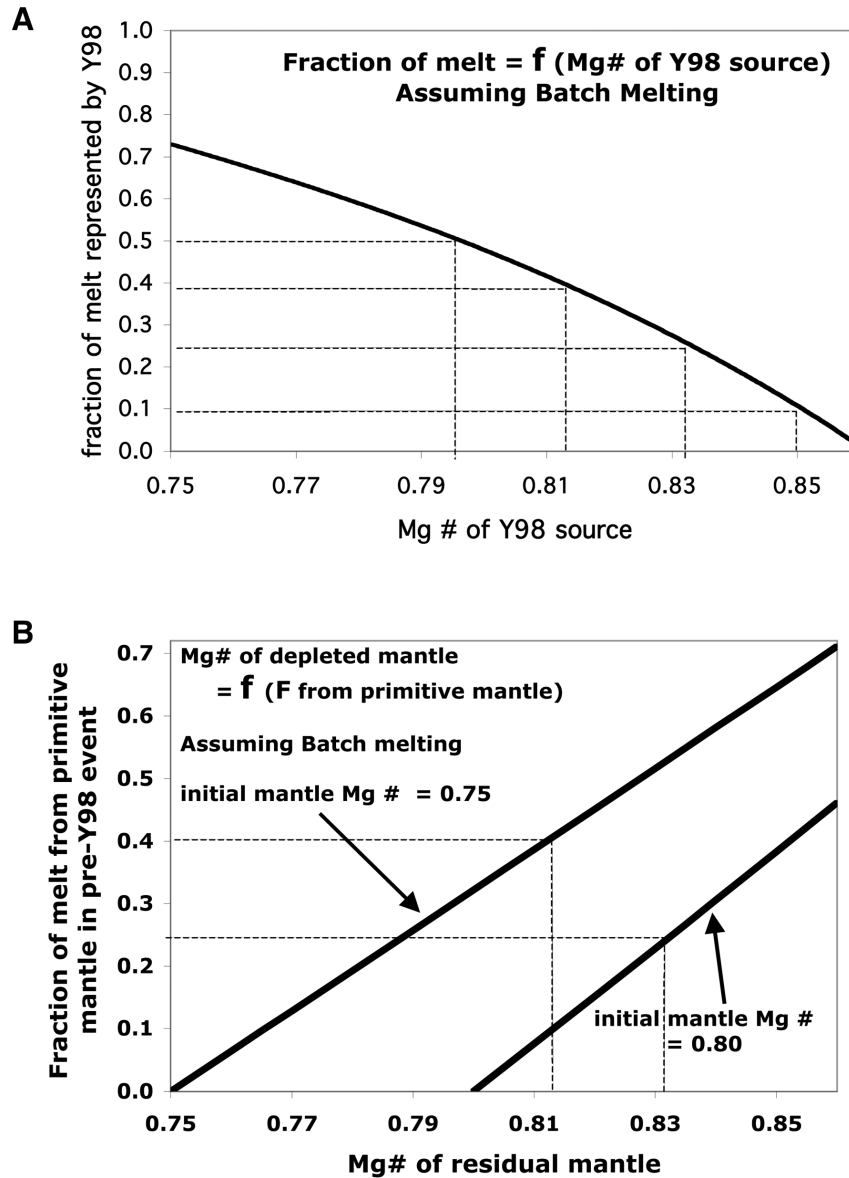


Fig. 5. a) The plot of the fraction of melt as a function of the assumed initial Mg# of the source based on the batch melting equation (Shaw 1970): $C_m/C_0 = 1/(F - D_i F + D_i)$; where C_m = concentration of element i in the melt, C_0 = concentration of element i in the bulk rock; D_i is the mineral/melt partition coefficient for element i , and F is the fraction of melt. The curve is solved for $F = (D_{Fe} - (A \times D_{Mg})) / (D_{Fe} - (A \times D_{Mg}) + A - 1)$; where $A = (Mg/Fe)_{melt} / (Mg/Fe)_{mantle} = (Mg\#_{melt} / (1 - Mg\#_{melt})) / (Mg\#_{mantle} / (1 - Mg\#_{mantle}))$, $D_{Fe} = 0.82$, $D_{Mg} = 2.6$ (Beattie et al. 1991; Jones 1995), $Mg\#_{melt} = 0.67$, and $Mg\#_{mantle} = 0.75$ to 0.86 (D values and Mg#s are molar). Dashed lines are tie-lines connecting values of F and Mg# discussed in text. b) The plot of the Mg# of the residuum in depleted mantle source as a function of the fraction of melt removed from the primitive mantle in a depletion event prior to the Y-980459 magmatism, assuming batch melting. Curves are solved for $F = (1 - B) / ((B \times (1/D_{Mg}) - 1) - ((1/D_{Fe}) + 1))$; where $B = (Mg/Fe)_{residuum} / (Mg/Fe)_0 = (Mg\#_{residuum} / (1 - Mg\#_{residuum})) / (Mg\#_0 / (1 - Mg\#_0))$, $Mg\#_0 = 0.75$ (upper curve), 0.80 (lower curve) $Mg\#_{residuum} = 0.75$ to 0.86 . Dashed lines are tie-lines connecting values of F and Mg# discussed in text.

mantle with an Mg# of 0.75, the lowest estimated primitive, or bulk, value. To achieve an Mg# of 0.86 in the residual mantle in the first depletion event (corresponding, in Fig. 5a, to infinitesimal melting of the depleted mantle, or Y-980459 source, in the second event) requires approximately 70% melting of the primitive mantle. If the fraction of melt is apportioned equally between the two partial melting events, then approximately 40% melting is required in each event.

From Fig. 5b, we see that the first 40% melting event results in an Mg# of 0.81 for the residual depleted mantle (or Y-980459 source). From Fig. 5a, we see that another 40% melting is required to produce the Y-980459 melt from a mantle source with an Mg# = 0.81. Such high melt fractions would have significant consequences for the thermal state of the Martian interior as discussed below.

A possible mitigating factor allowing for lower degrees

of partial melting is to consider a higher Mg# for the primitive or bulk Martian mantle. Based on high-pressure experimental evidence (Agee and Draper 2004, 2005), an Mg# of 0.80 for the primitive Martian mantle has been proposed (Borg and Draper 2003; Draper et al. 2005). The effect that a higher Mg# for the primitive Martian mantle has on the amount of partial melting necessary is illustrated by comparing Fig. 5a to the curve in the lower right side of Fig. 5b, which assumes an Mg# of 0.80 for the Martian mantle. In this case, if we apportion the fraction of melt equally between the two partial melting events, then approximately 25% melting during both the initial depletion event and the second event that produced Y-980459 is required, with the first depletion event resulting in an Mg# of 0.83 for the Y-980459 source prior to the Y-980459 melt production and extraction. This is much lower than the 40% melt considered above, but is still high. The extreme thermal constraints which such a large shift in the Mg# from primitive mantle to Y-980459 being accomplished by high degrees of partial melting places on the Martian mantle requires us to consider a lower percentage of melt during Y-980459 production with depletion occurring via several previous episodes of magmatism. If instead of the two-stage model considered above, we postulate multiple episodes of melt production, then each individual episode would require a lower amount of melt. For example, three episodes of 15% melting would be required to generate the Y-980459 magma from an initial primitive Mg# of 0.80.

Mineralogy of the Y-980459 Protolith

Given the compositions of the Y-980459 melt and the residual olivines and pyroxenes, we can estimate the modal composition of the Y-980459 source prior to the extraction of the Y-980459 melt for various assumed melt fractions (Table 4). The estimated composition and mode of the Y-980459 source is sensitive to the assumed olivine to pyroxene ratio in the residuum. The phases on the liquidus at multiple saturation tell us the minerals present in the residuum, but not their relative abundances. Given the high degree of melting inferred, whether it was prior to or during the Y-980459 magmatic episode, it is likely that pyroxene was depleted preferentially to olivine, so a high olivine-to-pyroxene ratio is likely for the residuum. For 40% melt with an Mg# of 0.64 in equilibrium with olivine and orthopyroxene with Mg# of 0.86 and a 5:1 olivine-to-pyroxene ratio in the residuum, the calculated mode for the Y-980459 source is approximately 58% olivine, 24% orthopyroxene, 10% clinopyroxene, 7% plagioclase, and less than 1% spinel, apatite, and troilite. A 1:1 olivine-to-pyroxene ratio for the residuum results in substantially lower olivine and higher orthopyroxene. A melt fraction of 10% gives a substantially higher olivine fraction and much lower plagioclase content.

IMPLICATIONS FOR THE THERMAL STATE OF THE MARTIAN MANTLE

We use the mantle convection and melt production model of Kiefer (2003) to assess the implications of these results for the thermal state of the interior of Mars. We summarize the numerical model briefly here; full details are provided in Kiefer (2003). This model simulates hot, rising mantle plumes in spherical axisymmetric geometry. Axisymmetric geometry implies that the rising plume is approximately cylindrical in cross-section. This approximation has been applied to many studies of mantle plumes, such as Hawai'i on Earth (e.g., Watson and McKenzie 1991; Ribe and Christensen 1999; van Hunen and Zhong 2003), and is a first-order description of the structure beneath Hawai'i imaged by seismic tomography (Lei and Zhao 2006). The numerical model includes a number of important physical effects, including the presence of a stiff, high-viscosity lithosphere and the partitioning of radioactive heating between the mantle and crust. The partial differential equations for the conservation of mass, momentum, and energy are solved using finite element methods (Kellogg and King 1997) on a hemispherical numerical grid with 128 elements in the vertical dimension and 256 elements in the horizontal, corresponding to a resolution of 13×21 km. The radioactive heating rates considered here result in thermal Rayleigh numbers (a measure of convective vigor) of less than 5×10^6 , and this numerical grid fully resolves the associated thermal structures.

The basic mantle thermal structure in our models resembles that shown in Fig. 3 of Kiefer (2003). Each simulation was integrated for a sufficiently long period of time, typically 1–2 billion years, to ensure that it reached secular equilibrium with the imposed radioactive heating rate. Secular equilibrium was assessed by the absence of a long-term secular trend in the mean mantle temperature. Once secular equilibrium is reached, there still may be short-term oscillations in the mantle temperature due to the development of boundary-layer instabilities, which are a natural part of vigorously convecting systems (e.g., Figs. 1 and 5 of Kiefer 2003). We have therefore integrated the numerical simulations for at least 30,000 time-steps, equivalent to several billion years, after secular equilibrium was achieved to fully assess the range of thermal states achieved by each set of model parameters.

Melting calculations follow the conceptual approach of Watson and McKenzie (1991) of tracing streamlines through the melting zone and calculating the differential melt production at each point, although the calculations are done with a Mars-specific melting model. The original models of Kiefer (2003) used experimental measurements on Mars analog mantle compositions by Bertka and Holloway (1994) and by Agee and Draper (2004) to define the mantle solidus. In the new results described here, we enhance the original

Table 4. The results of modal calculations for the Y-980459 source for different assumed fraction of melt and olivine: pyroxene ratios in the source residuum.

Melt:ol:pxn	40:50:10	40:30:30	10:75:15	10:45:45
Mg#	0.81	0.80	0.85	0.84
SiO ₂	45.0	48.8	42.9	48.6
TiO ₂	0.2	0.2	0.1	0.1
Al ₂ O ₃	2.3	2.4	0.7	0.8
Cr ₂ O ₃	0.6	0.6	0.5	0.5
FeO	14.4	13.6	13.3	12.1
NiO	0.04	0.04	0.05	0.04
MnO	0.4	0.4	0.4	0.4
MgO	33.6	30.2	40.9	35.9
CaO	3.0	3.2	1.1	1.4
Na ₂ O	0.3	0.3	0.1	0.1
K ₂ O	0.01	0.01	0.01	0.00
P ₂ O ₅	0.14	0.14	0.06	0.05
FeS	0.05	0.05	0.01	0.01
Total	100	100	100	100
Ol	58.0	35.3	77.9	46.8
Opx	23.6	44.8	15.2	43.6
Cpx	10.2	11.3	3.9	5.9
Plag	7.4 (An ₆₈)	7.6 (An ₆₉)	2.2 (An ₆₈)	2.8 (An ₇₁)
Apatite	0.1	0.1	0.0	0.04
Oxide	0.8	0.8	0.8	0.8
Troilite	0.05	0.05	0.01	0.01
Total	100	100	100	100

model by including a new solidus appropriate for the Y-980459 composition and by considering localized thinning of the lithosphere over the mantle plume.

Given a particular choice of mantle solidus, three model parameters control the ability of an ascending mantle plume to generate melt by adiabatic decompression melting: the temperature at the base of the mantle, the amount of radioactivity in the mantle, and the thickness of the lithosphere. The temperature at the base of the mantle sets the initial temperature of the rising mantle plume; the hotter the initial plume temperature, the more likely it is to partially melt as it approaches the surface of Mars. The amount of mantle radioactivity affects the temperature of the mantle and thus the amount of conductive cooling experienced by a rising mantle plume. Increasing the mantle radioactive heating rate increases the mean mantle temperature and decreases the rate of plume cooling, thus increasing the magma production rate. The maximum ascent of the mantle plume and thus the extent of adiabatic decompression melting are controlled by the lithosphere's thickness. Thinning the lithosphere increases the amount of magma produced by the rising mantle plume.

The models of Kiefer (2003) simulated the lithosphere as a high-viscosity layer of specified, laterally constant thickness. The viscosity decreased exponentially with depth below the surface, with the viscosity at the surface equal to 10⁵ times the mantle viscosity. The thickness of the high-viscosity layer was varied between 170 and 250 km, with a nominal value in most models of 212 km. This produced a

heat flux out of the mantle that is consistent with the mantle heat flux estimated by McGovern et al. (2002, 2004) on the basis of gravity modeling of the large volcanoes in Tharsis and Elysium. The gravity results constrain the lithospheric thickness averaged over regions that are at least 1000–2000 km across. However, heating of the lithosphere by a mantle plume is likely to lead to localized thinning of the lithosphere. This allows the plume to ascend closer to the surface, which may be quite important in calculating the expected magma production from the plume.

Motivated by the determination of multiple saturation of the Y-980459 source composition near 12 kbar (approximately 100 km depth), our new plume simulations consider the effects of thinner lithosphere along the upwelling plume axis. For simplicity, we assume that the lithospheric thickness varies in a roughly Gaussian manner with distance from the plume axis. The lithospheric thickness is a minimum along the plume axis and increases with distance from the upwelling. We interpret the Y-980459 multiple saturation pressure as the minimum pressure at which melt and solid were in equilibrium and thus as evidence for the minimum depth of melting experienced by this magma body. The minimum lithospheric thickness in the numerical models is thus assumed to be 106 km on the plume axis. Beyond 170 km from the plume axis, the lithosphere has a fixed thickness of 212 km, as in the earlier models of Kiefer (2003). The specific model lithosphere thicknesses of 106 and 212 km are set by the structure of the finite element grid used in the modeling;

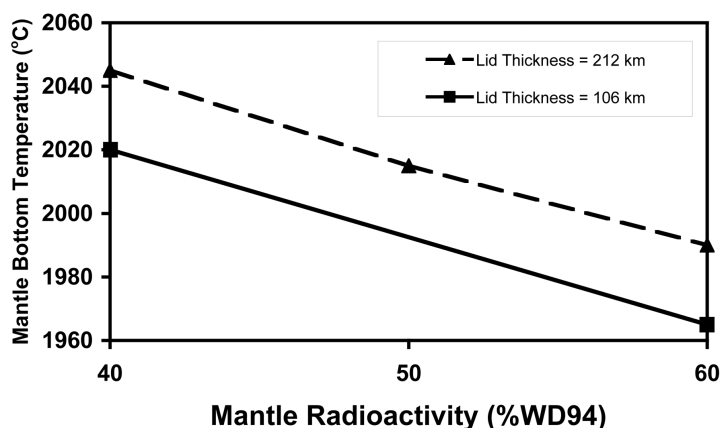


Fig. 6. The minimum required temperature at the base of the convecting mantle to permit melting of Y-980459 in an ascending mantle plume. Results are shown as a function of the assumed mantle radioactivity content, expressed as a fraction of the Wänke and Dreibus (1994) total present-day radioactivity. The solid line and squares are results for a lithosphere thickness of 106 km on the central plume axis. The dashed line and triangles are for a lithosphere thickness of 212 km.

they should not be interpreted as constraining the actual lithosphere thickness to three significant figures. Although the specific shape of the imposed lithospheric thinning is somewhat arbitrary, it is consistent with previous mantle plume modeling (e.g., Ribe and Christensen 1999). A more realistic and computationally demanding approach would be to use a fully temperature-dependent viscosity law (e.g., Mei and Kohlstedt 2000) and allow the lithospheric thinning to develop self-consistently via coupling to the underlying thermal structure. Such models are currently in development. When completed, we will be able to compare the lithospheric thinning in the numerical modeling with that inferred from the Y-980459 multiple saturation pressure as an additional test of the numerical models.

We use the chemical composition model of Wänke and Dreibus (1994) for the overall abundance of radioactive elements in the silicate portion of Mars, which we then partition between the crust and mantle. In these models, the crust is assumed to be 53 km in thickness (the top four rows of the finite element grid), consistent with the results of Wieczorek and Zuber (2004). Kiefer (2003) found that retaining between 40 and 60% of the total radioactivity within the mantle, with the remainder in the crust, is consistent with the overall range of geologic, geochemical, and geophysical constraints on the Martian mantle. We consider the same range of mantle radioactivity here.

Given these model assumptions, we vary the assumed temperature at the base of the mantle, T_{base} , in order to determine the minimum bottom temperature required to reach the observed Y-980459 melting curve shown in Fig. 4. Our numerical results are summarized in Fig. 6, which shows the minimum temperature at the base of the convecting layer that is necessary to permit the ascending plume to produce the observed Y-980459 melt by adiabatic decompression. If the convecting layer extends throughout the entire mantle, T_{base} corresponds to the temperature at the core-mantle boundary.

Alternatively, if there is chemical stratification in the lower mantle of Mars, then T_{base} corresponds to the temperature at the interface between the convecting upper mantle and the deeper, chemically distinct layer. For a lithosphere thickness of 106 km (solid line and squares), as determined from the multiple saturation pressure, T_{base} must be at least 1965–2020 °C, depending on the amount of radioactive heating in the mantle. Larger values of mantle radioactive heating result in a higher mean mantle temperature and thus less cooling of the plume, and thus permit melting to occur with smaller values of T_{base} . For comparison, the dashed line and triangles in Fig. 6 are the required T_{base} values if the lithosphere is 212 km thick, as in the earlier work of Kiefer (2003). The thinner lithosphere modeled here permits a greater amount of adiabatic decompression and thus melting initiates at a somewhat lower temperature.

These temperatures do not include the effects of latent heat of melting and thus are minimum values for T_{base} . The required amount of energy for the latent heat of melting depends on the actual melt fraction represented by the Y-980459 magma. The temperatures listed above must be increased by $\delta T = X \times (\Delta H_m / C_p)$, where X is the melt fraction, ΔH_m is the latent heat of melting, and C_p is the specific heat. For the Y-980459 composition, plausible values are $\Delta H_m = 6.4 \times 10^5 \text{ J kg}^{-1}$ and $C_p = 1200 \text{ J K}^{-1} \text{ kg}^{-1}$ (Navrotsky 1995). For example, if Y-980459 represents a 5% melt fraction, then the values of T_{base} required to produce the observed melt must be increased by $\delta T = 27 \text{ °C}$. If the melt fraction is higher, as considered above in the discussion of Fig. 5, then the required value of δT will also be higher.

For comparison, Kiefer (2003) found that a minimum T_{base} of 1760–1790 °C is required to explain the recent magma production rate on Mars using the Bertka and Holloway (1994) solidus. The much higher values of T_{base} reported here reflect the very high melting temperature of the Y-980459 composition and thus place significant new constraints on the

thermal structure of the Martian mantle and core. The plausibility of the required values of T_{base} can be assessed by comparison with the results of Williams and Nimmo (2004), who used a parameterized convection model to calculate the thermal history of the Martian core. Their results are expressed in terms of potential temperature, which removes the effect of the adiabatic temperature gradient from the results. Thus, to compare their results to our melting calculations, we have added 306 °C to their results (mantle plus crust thickness 1698 km, mean adiabatic gradient 0.18 K/km [Kiefer 2003]). Y-980459 has an estimated igneous formation age of 472 ± 47 Ma (Shih et al. 2005), and its mantle plume required roughly 100–200 million years to ascend through the mantle. Thus, our results for T_{base} should be compared with Williams and Nimmo's (2004) results for 600–700 million years ago. Williams and Nimmo estimated a core temperature at that time of approximately 1870 °C, which is 95 °C less than our minimum required value of T_{base} and several hundred degrees less than estimates involving high degrees of partial melting. In contrast to the conclusions based on the partial melting models above, these thermal considerations suggest that the final Y-980459 melting event was probably a low melt fraction event. If this is the case, then the earlier depletion event(s) must have involved cumulatively very high percentage melting and probably involved more than one cycle of melting prior to the Y-980459 event. But even with a very low melt fraction, the minimum required value for T_{base} is significantly higher than the Williams and Nimmo (2004) estimate of the core temperature.

There are several possible resolutions to this discrepancy. First, if the Martian core lacks sulfur and is presently solidifying, the release of latent heat would buffer the core temperature near 2030 °C, but this is inconsistent with the absence of a present-day magnetic field (Williams and Nimmo 2004). Second, if the mantle viscosity is higher than assumed in these models, the convective heat flux would be reduced and the core would be hotter than calculated. (Conversely, decreasing the mantle viscosity would result in a cooler core and would magnify the mismatch between our results and those of Williams and Nimmo). However, it is not clear how the required viscosity increase could be achieved.

A third possibility, suggested by Jones (2003) and Kiefer (2003), is that there may be a dense, chemically distinct layer at the base of the Martian mantle that is enriched in radioactive elements. The mantle source region for the shergottite meteorites has long been recognized as being depleted in aluminum (Longhi 1991; Treiman 2003). Borg and Draper (2003) showed that the super-chondritic CaO/ Al_2O_3 ratio in the Martian meteorites cannot be explained by segregation of crustal minerals and is best explained by crystallization of garnet in an early Martian magma ocean. Because of its high density, garnet crystallizing in a magma ocean would naturally sink to the base of the mantle (Elkins-

Tanton et al. 2003), and the high density would inhibit later convective remixing of the deep layer into the overlying mantle (Kellogg and King 1993). Another phase that would plausibly be present in this deep layer is Ca perovskite (Bertka and Fei 1997), which can contain significant U and Th (Taura et al. 2001; Corgne and Wood 2002). Thus, if such a chemically distinct layer exists at the base of the Martian mantle, radioactive heating within the layer could contribute to the high value of T_{base} required to produce the Y-980459 melts and might even accommodate the high inferred melt fraction. Elkins-Tanton et al. (2005a, 2005b) discuss how such a deep, radioactively enriched layer may influence generation of a magnetic field and initial crustal formation on early Mars. However, detailed calculations of how such a deep layer might affect recent volcanism such as melting of the Y-980459 parent composition have not yet been performed.

A fourth possible solution to this discrepancy is if the water content of the Y-980459 mantle source is high enough to significantly lower the melting point. Water contents in the 1000–5000 ppm range can lower the solidus temperature by hundreds of degrees (e.g., Wyllie 1979). No published direct measurements of the water content of Y-980459 are available at this time. However, as is generally the case with Martian meteorites there is no petrographic evidence, such as the existence of hydrated minerals, for significant water in Y-980459. The water content of Martian meteorites is generally low—on the order of 200 ppm (e.g., Leshin et al. 1996). It is possible, of course, that their parent magmas were wetter than this amount and that the Martian lavas that we have sampled are universally outgassed to a high degree. The water content of the Martian mantle is the subject of some controversy. Estimates range from 100 to 200 ppm (Jones 2004) up to 2 wt% or more (Dann et al. 2001). However, increasing the mantle's water content will also decrease the mantle's viscosity (e.g., Mei and Kohlstedt 2000). This in turn will lead to more vigorous convection and a faster rate of heat loss over geologic time and thus a cooler mantle. This will at least partially and perhaps fully offset the effects of the lower melting temperature. Numerical simulations are currently in development that will address this trade-off in a self-consistent manner.

SUMMARY AND CONCLUSIONS

The results of our experiments give strong evidence that Y-980459 represents its parental melt. The first-formed experimental olivines have an Mg# that is identical to the cores of the natural phenocrysts. Furthermore, the implication is strong that Y-980459 is a primary melt of the Martian mantle. The liquidus becomes saturated with both olivine and low-Ca pyroxene at approximately 12 kbar and 1540 °C. This multiple saturation point is unique for the Y-980459 composition with olivine only on the liquidus at lower

pressures and pyroxene only on the liquidus at higher pressures.

The conditions and mineralogy of multiple saturation provide insight into the mantle source for Y-980459. The depth of magma separation from the mantle is at about 100 km, indicating the depth of the lithosphere below the Y-980459 eruption area. The high Mg content of the source residuum suggests a large amount of magma extraction previous to the Y-980459 eruption, and perhaps a large melt fraction for Y-980459 itself. This conclusion is consistent with inferences made from Rb/Sr systematics.

The high temperature indicated for the upper Martian mantle and possibly high melt fraction for the Y-980459 magmatic event, when factored into convective models of the Martian interior, suggest a significantly higher temperature at the core-mantle boundary than previously estimated. The higher inferred interior temperatures require revision of our views of the Martian interior.

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REFERENCES

- Agee C. B. and Draper D. S. 2004. Experimental constraints on the origin of Martian meteorites and the composition of the Martian mantle. *Earth and Planetary Science Letters* 224:415–429.
- Agee C. B. and Draper D. S. 2005. High-pressure melting of H chondrite: A match for the Martian basalt source mantle (abstract #1434). 36th Lunar and Planetary Science Conference. CD-ROM.
- Barrat J.-A., Jambon A., Bohn M., Gillet Ph., Sautter V., Gopel C., Lesourd M., and Keller F. 2002. Petrology and chemistry of the picritic shergottite Northwest Africa 1068 (NWA 1068). *Geochimica et Cosmochimica Acta* 66:3505–3518.
- Beattie P., Clifford F., and Russell D. 1991. Partition coefficients for olivine-melt and orthopyroxene-melt systems. *Contributions to Mineralogy and Petrology* 109:212–224.
- Bertka C. M. and Fei Y. 1997. Mineralogy of the Martian interior up to core-mantle boundary pressures. *Journal of Geophysical Research* 102:5251–5264.
- Bertka C. M. and Holloway J. R. 1994. Anhydrous partial melting of an iron-rich mantle I: Subsolvus phase assemblages and partial melting phase relationships at 10 to 30 kbar. *Contributions to Mineralogy and Petrology* 115:313–322.
- Borg L. E. and Draper D. S. 2003. A petrogenetic model for the origin and compositional variation of the Martian basaltic meteorites. *Meteoritics & Planetary Science* 38:1713–1731.
- Borg L. E., Nyquist L. E., Taylor L. A., Wiesmann H., and Shih C.-Y. 1997. Constraints on Martian differentiation processes from Rb-Sr and Sm-Nd isotopic analyses of the basaltic shergottite QUE 94201. *Geochimica et Cosmochimica Acta* 61:4915–4931.
- Borg L. E., Nyquist L. E., Wiesmann H., Shih C.-Y., and Reese Y. 2003. The age of Dar al Gani 476 and the differentiation history of the Martian meteorites inferred from their radiogenic isotope systematics. *Geochimica et Cosmochimica Acta* 67:3519–3536.
- Boyd F. R. and England J. C. 1963. Effect of pressure on the melting of diopside, $\text{CaMgSi}_2\text{O}_6$ and albite, $\text{NaAlSi}_3\text{O}_8$ in the range up to 50 kb. *Journal of Geophysical Research* 68:311–323.
- Corgne A. and Wood B. J. 2002. CaSiO_3 and CaTiO_3 perovskite-melt partitioning of trace elements: Implications for gross mantle differentiation. *Geophysical Research Letters*, doi:10.1029/2001GL014398.
- Dann J. C., Holzheid A. H., Grove T. L., and McSween H. Y., Jr. 2001. Phase equilibria of the Shergotty meteorite: Constraints on pre-eruptive H_2O contents of Martian magmas and fractional crystallization under hydrous conditions. *Meteoritics & Planetary Science* 36:793–806.
- Draper D. S., Borg L. E., and Agee C. B. 2005. Crystallization of a Martian magma ocean and the formation of shergottite source regions (abstract #1492). 36th Lunar Planetary Science Conference. CD-ROM.
- Dreibus G. and Wänke H. 1987. Volatiles on Earth and Mars: A comparison. *Icarus* 71:225–240.
- Dreibus G., Haubold R., Huisl W., and Spettel B. 2003. Comparison of the chemistry of Yamato-980459 with DaG 476 and SaU 005 (abstract). International Symposium, Evolution of Solar System Materials: A New Perspective from Antarctic Meteorites. pp. 19–20.
- Dunn T. 1993. The piston-cylinder apparatus. In *Experiments at high pressure and applications to the Earth's mantle*, edited by Luth R. W. Québec, Ontario: Mineralogical Association of Canada. pp. 39–94.
- Elkins-Tanton L. T., Zaranek S. E., Parmentier E. M., and Hess P. C. 2005a. Early magnetic field and magmatic activity on Mars from magma ocean cumulate overturn. *Earth and Planetary Science Letters* 236:1–12.
- Elkins-Tanton L. T., Hess P. C., and Parmentier E. M. 2005b. Possible formation of ancient crust on Mars through magma ocean processes. *Journal of Geophysical Research*, doi:10.1029/2005JE002480.
- Elkins-Tanton L. T., Parmentier E. M., and Hess P. C. 2003. Magma ocean fractional crystallization and cumulate overturn in terrestrial planets: Implications for Mars. *Meteoritics & Planetary Science* 38:1753–1771.
- Ghiorso M. S. and Sack R. O. 1995. Chemical mass transfer in magmatic processes IV. A revised and internally consistent thermodynamic model for the interpolation and extrapolation of liquid-solid equilibria in magmatic systems at elevated temperatures and pressures. *Contributions to Mineralogy and Petrology* 119:197–212.
- Goodrich C. A. 2003. Petrogenesis of olivine-phyric shergottites Sayh al Uhaymir 005 and Elephant Moraine 79001 lithology A. *Geochimica et Cosmochimica Acta* 67:3735–3771.
- Goodrich C. A., Herd C. D. K., and Taylor L. A. 2003. Spinel and oxygen fugacity in olivine-phyric and lherzolitic shergottites. *Meteoritics & Planetary Science* 38:1773–1792.
- Greshake A., Fritz J., and Stöffler D. 2004. Petrology and metamorphism of the olivine-phyric shergottite Yamato-980459: Evidence for a two-stage cooling and single-stage ejection history. *Geochimica et Cosmochimica Acta* 68:2359–2377.
- Herd C. D. K. 2003. The oxygen fugacity of olivine-phyric Martian basalts and the components within the mantle and crust of Mars. *Meteoritics & Planetary Science* 38:1793–1805.
- Herd C. D. K. Forthcoming. Insights into the redox history of the NWA 1068/1110 Martian basalt from mineral equilibria and

- vanadium oxybarometry. *Meteoritics & Planetary Science*.
- Herd C. D. K., Papike J. J., and Brearley A. J. 2001. Oxygen fugacity of Martian basalts from electron microprobe oxygen and TEM-EELS analyses of Fe-Ti oxides. *American Mineralogist* 86: 1015–1024.
- Holloway J. R., Pan V., and Gudmundsson G. 1992. High-pressure fluid absent melting experiments in the presence of graphite: Oxygen fugacity, ferric/ferrous ratio and dissolved CO₂. *European Journal of Mineralogy* 4:105–114.
- Jones J. H. 1995. Experimental trace element partitioning. In *Rock physics and phase relations: A handbook of physical constants*, edited by Ahrens T. J. Washington, D.C.: American Geophysical Union. pp. 73–104.
- Jones J. H. 2003. Constraints on the structure of the Martian interior from the chemical and isotopic systematics of the SNC meteorites. *Meteoritics & Planetary Science* 38:1807–1814.
- Jones J. H. 2004. The edge of wetness: The case for dry magmatism on Mars (abstract # 1798). 35th Lunar and Planetary Science Conference. CD-ROM.
- Kellogg L. H. and King S. D. 1997. The effect of temperature dependent viscosity on the structure of new plumes in the mantle: Results of a finite element model in a spherical, axisymmetric shell. *Earth and Planetary Science Letters* 148:13–26.
- Kellogg L. H. and King S. D. 1993. Effect of mantle plumes on the growth of the D double-prime region by reaction between the core and mantle. *Geophysical Research Letters* 20:379–382.
- Kilinc A., Carmichael I. S. E., Rivers M. L., and Sack R. O. 1983. The ferric-ferrous ratio of natural silicate liquids equilibrated in air. *Contributions to Mineralogy and Petrology* 83:136–140.
- Kiefer W. S. 2003. Melting in the Martian mantle: Shergottite formation and implications for present-day mantle convection on Mars. *Meteoritics & Planetary Science* 38:1815–1832.
- Koizumi E., Mikouchi T., Monkawa A., and Miyamoto M. 2003. The origin of olivine megacryst in the Dar al Gani 476 basaltic shergottite and its paired meteorites: Verification of three models (abstract). International Symposium, Evolution of Solar System Materials: A New Perspective from Antarctic Meteorites. pp. 56–57.
- Kress V. C. and Carmichael I. S. E. 1988. Stoichiometry of the iron oxidation reaction in silicate melts. *American Mineralogist* 73: 1267–1274.
- Lei J. and Zhao D. 2006. A new insight into the Hawai'ian plume. *Earth and Planetary Science Letters* 241:438–453.
- Leshin L. A., Epstein S., and Stolper E. M. 1996. Hydrogen isotope geochemistry of SNC meteorites. *Geochimica et Cosmochimica Acta* 60:2635–2650.
- Longhi J. 1991. Complex magmatic processes on Mars: Inferences from the SNC meteorites. Proceedings, 21st Lunar and Planetary Science Conference. pp. 695–709.
- Longhi J., Knittle E., Holloway J. R., and Wänke H. 1992. The bulk composition, mineralogy and internal structure of Mars. In *Mars*, edited by Kiefer H. H., Jakosky B. M., Snyder C. W., and Matthews M. S. Tucson, Arizona: The University of Arizona Press. pp. 184–208.
- McCanta M., Rutherford M. J., and Jones J. H. 2004. An experimental study of rare earth element partitioning between a shergottite melt and pigeonite: Implications for the oxygen fugacity of the Martian interior. *Geochimica et Cosmochimica Acta* 68:1943–1952.
- McDade P., Wood B. J., van Westrenen W., Brooker R., Gudmundsson G., Soular H., Najoka J., and Blundy J. 2002. Pressure corrections for a selection of piston cylinder cell assemblies. *Mineralogical Magazine* 66:1021–1028.
- McGovern P. J., Solomon S. C., Smith D. E., Zuber M. T., Simons M., Wieczorek M. A., Phillips R. T., Neuman G. A., Aharonson O., and Head J. W. 2002. Localized gravity/topography admittance and correlation spectra on Mars: Implications for regional and global evolution. *Journal of Geophysical Research*, doi:10.1029/2002JE001854.
- McGovern P. J., Solomon S. C., Smith D. E., Zuber M. T., Simons M., Wieczorek M. A., Phillips R. T., Neuman G. A., Aharonson O., and Head J. W. 2004. A correction to “Localized gravity/topography admittance and correlation spectra on Mars: Implications for regional and global evolution.” *Journal of Geophysical Research*, doi: 10.1029/2004JE002286.
- McKay G., Le L., Schwandt C., Mikouchi T., Koizumi E., and Jones J. 2004. Yamato-980459: The most primitive shergottite (abstract #2154)? 35th Lunar and Planetary Science Conference. CD-ROM.
- Mei S. and Kohlstedt D. L. 2000. Influence of water on plastic deformation of olivine aggregates 1. Diffusion creep regime. *Journal of Geophysical Research* 105:21,457–21,469.
- Mikouchi T., Koizumi E., McKay G., Monkawa A., Ueda Y., and Miyamoto M. 2003. Mineralogy and petrology of the Yamato-980459 Martian meteorite: A new shergottite-related rock (abstract). International Symposium, Evolution of Solar System Materials: A New Perspective from Antarctic Meteorites. pp. 82–83.
- Misawa K. 2003. The Yamato-980459 shergottite consortium (abstract). International Symposium, Evolution of Solar System Materials: A New Perspective from Antarctic Meteorites. pp. 84–85.
- Monders A. G., Medard E., and Grove T. L. 2006. Basaltic lavas at Gusev crater revisited (abstract #1834). 37th Lunar and Planetary Science Conference. CD-ROM.
- Navrotsky A. 1995. Thermodynamic properties of minerals. In *Mineral physics and crystallography: A handbook of physical constants*, edited by Ahrens T. J. Washington, D.C.: American Geophysical Union. pp. 18–28.
- Neuman G. A., Zuber M. T., Wieczorek M. A., McGovern P. J., Lemoine F. G., and Smith D. E. 2004. Crustal structure of Mars from gravity and topography. *Journal of Geophysical Research*, doi:10.1029/2004JE002262.
- Norris J. R. and Herd C. D. K. 2006. The Yamato-980459 liquidus at 10 to 20 kilobars (abstract #1787). 37th Lunar and Planetary Science Conference. CD-ROM.
- Redmond H. L. and King S. D. 2004. A numerical study of a mantle plume beneath the Tharsis rise: Reconciling dynamic uplift and lithospheric support models. *Journal of Geophysical Research*, doi:10.1029/2003JE002228.
- Ribe N. M. and Christensen U. R. 1999. The dynamical origin of Hawai'ian volcanism. *Earth and Planetary Science Letters* 171: 517–531.
- Shaw D. M. 1970. Trace element fractionation during anatexis. *Geochimica et Cosmochimica Acta* 68:2359–2377.
- Shih C.-Y., Nyquist L. E., Wiesmann H., Reese Y., and Misawa K. 2005. Rb-Sr and Sm-Nd dating of olivine-phyric shergottite Yamato-980459: Petrogenesis of depleted shergottites. *Antarctic Meteorite Research* 18:46–65.
- Shirai N. and Ebihara M. 2004. Chemical characteristics of an olivine-phyric shergottites, Yamato-980459 (abstract #1511). 35th Lunar and Planetary Science Conference. CD-ROM.
- Taura H., Yurimoto H., Kato T., and Sueno S. 2001. Trace element partitioning between silicate perovskites and ultracalcic melts. *Physics of the Earth and Planetary Interiors* 124:25–32.
- Treiman A. H. 2003. Chemical compositions of Martian basalts (shergottites): Some inferences on basalt formation, mantle metasomatism, and differentiation in Mars. *Meteoritics & Planetary Science* 38:1849–1864.
- Toplis M. J. 2005. The thermodynamics of iron and magnesium partitioning between olivine and liquid: Criteria for assessing and predicting equilibrium in natural and

- experimental systems. *Contributions to Mineralogy and Petrology* 149:22–39.
- van Hunen J. and Zhong S. 2003. New insight in the Hawaiian plume swell dynamics from scaling laws. *Geophysical Research Letters*, doi:10.1029/2003GL017646.
- Wadhwa M. 2001. Redox state of Mars' upper mantle and crust from Eu anomalies in shergottite pyroxenes. *Science* 291:1527–1530.
- Wänke H. and Dreibus G. 1994. Chemistry and accretion history of Mars. *Philosophical Transactions of the Royal Society of London A* 348:285–293.
- Watson E. B., Wark D. A., Price J. D., and van Orman J. A. 2002. Mapping the thermal structure of solid-media pressure assemblies. *Contributions to Mineralogy and Petrology*, doi: 10.1007/s00410-001-0327-4.
- Watson S. and McKenzie D. 1991. Melt generation by plumes: A study of Hawaiian volcanism. *Journal of Petrology* 32:501–537.
- Wieczorek M. A. and Zuber M. T. 2004. Thickness of the Martian crust: Improved constraints from geoid-to-topography ratios. *Journal of Geophysical Research*, doi:10.1029/2003JE002153.
- Williams D. W. and Kennedy J. C. 1969. Melting curve of diopside to 50 kilobars. *Journal of Geophysical Research* 74:4359–4366.
- Williams J.-P. and Nimmo F. 2004. Thermal evolution of the Martian core: Implications for an early dynamo. *Geology* 32:97–100.
- Wyllie P. J. 1979. Magmas and volatile components. *American Mineralogist* 64:469–500.
- Zuber M. T., Solomon S. C., Phillips R. T., Smith D. E., Tyler G. L., Aharonson O., Balmino G., Banerdt W. B., Head J. W., Johnson C. L., Lemoine F. G., McGovern P. J., Neuman G. A., Rowlands D. D., and Zhong S. 2000. Internal structure and early thermal evolution of Mars from Mars Global Surveyor topography and gravity. *Science* 287:1788–1793.

APPENDIX: PISTON CYLINDER CALIBRATION DATA

Temperature Calibration

A temperature profile for the piston cylinder assembly in the region of the sample charge was determined using the spinel growth-rate method of Watson et al. (2002). A furnace/pressure assembly identical to those used in the Y-980459 experiments was employed with a magnesia/alumina “sandwich” placed into the sample capsule in place of the usual oxide powder mixture. A graphite capsule was placed into the furnace assembly in its usual spot above the thermocouple. A magnesia bushing with the bottom end machined to fit inside the sample capsule was placed on top of the capsule so that the machined bottom end was inserted into the capsule. An alumina thermocouple sheath was inserted into the axial bore of the magnesia bushing. The magnesia bushing and thermocouple sheath extended 2 mm above the capsule. A solid magnesia bushing was placed above this to fill the remaining space inside the graphite furnace. As with the Y-980459 experiments, the pressure and temperature ramps were done “hot piston out,” with temperature ramped up over 35 min. The run was kept at 12 kbar and 1500 °C for 25 h, then quenched isobarically. The inner portion of the assembly was mounted in epoxy and carefully sectioned to its mid-plane. An optical image of the sectioned assembly is shown in Fig. A1.

The thickness of the spinel reaction layer was measured at various points along the alumina/magnesia interface from high magnification optical images. Temperatures were calculated from the spinel thickness using both polybaric-fit and isobaric-fit equations from Watson et al. (2002). For the isobaric calculation, the temperature was determined by a linear interpolation between the 10 kbar and 17 kbar equations.

Polybaric fit equation: $T = -B/(\ln((\Delta X^2)/A \times t) + 2.08 p^{1/2})$

Isobaric fit equation: $T = -B/(\ln(\Delta X^2)/A \times t)$

where A , B are fit parameters; $A = 8.58 \times 10^{11}$ (polybaric),

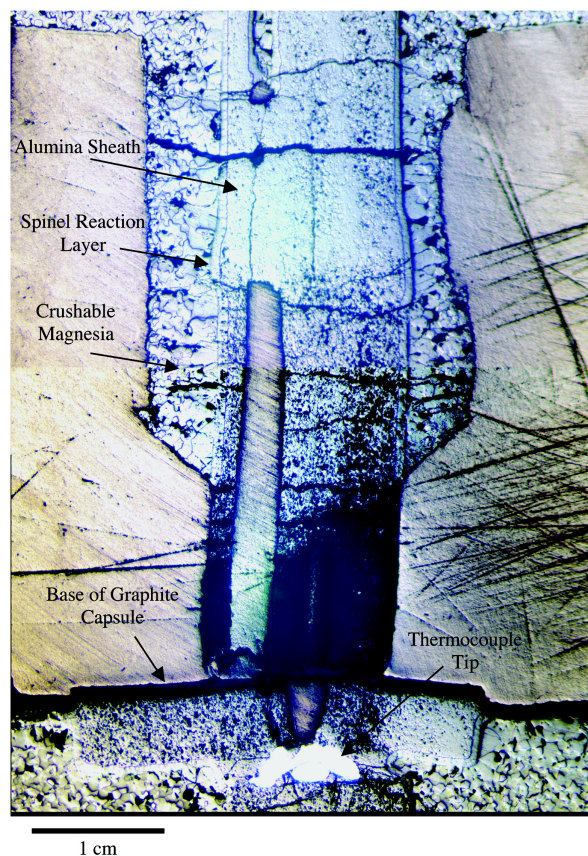


Fig. A1. A reflected light photomicrograph of the vertically sectioned run product from the temperature-profile experiment. Note spinel reaction layer between the alumina thermocouple sheath and the crushable magnesia inside the graphite capsule. The alumina sheath has pushed to the base of the graphite capsule, with some of the graphite having intruded into the bore that would normally house the thermocouple wire.

1.32×10^{11} (10 kbar isobaric), 1.28×10^{11} (17 kbar isobaric); $B = 48865$ (polybaric), 49202 (10 kbar isobaric), 50261 (17 kbar isobaric); T = temperature in K; ΔX = thickness of the spinel layer in μ ; t = time at temperature in seconds; and p = pressure in GPa. The calculated temperature profile is shown in Fig. A2.

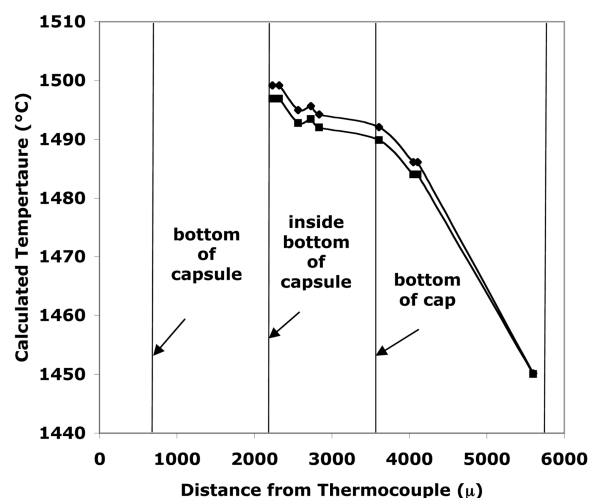


Fig. A2. The plot of temperature profile calculated from the thickness of the spinel layer in the temperature profile experiment using the spinel-growth method of Watson et al. (2002). The temperatures plotted as diamonds were calculated using the polybaric-fit equation. The temperatures plotted as squares were interpolated from calculations using the isobaric-fit equations for 10 and 17 kbar.

Pressure Calibration

The pressure was calibrated using the melting point of diopside (Boyd and England 1963; Williams and Kennedy 1969). A furnace/pressure assembly identical to those used in the Y-980459 experiments was employed. Synthetic diopside powder as either crystal or glass was loaded into graphite capsules. Temperature and pressure ramps were conducted in

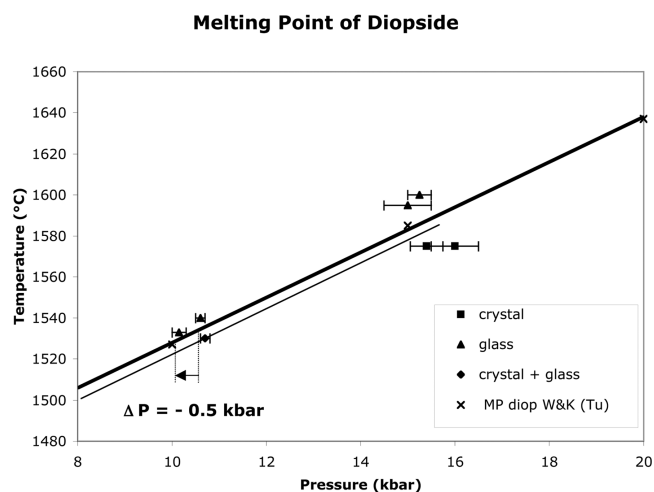


Fig. A3. The plot of results from diopside melting point determination experiments. The thick line is the diopside melting point curve from Williams and Kennedy (1969). The thin line is a fit to our data. There is a 0.5 ± 0.1 kbar offset between the two data sets.

a manner identical to that employed for the Y-980459 experiments. Run products were sectioned and mounted on glass slides and polished. The presence of crystal or glass was determined both visually and from backscattered electron images. The run temperatures and pressures and resulting phases are plotted in Fig. A3. Also plotted are the diopside melting point curve from Williams and Kennedy (1969) and melting point curve from our data. There is a 0.5 ± 0.1 kbar offset between the two data sets.