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Comment

Comment on “The origin of eucrites, diogenites, and olivine diogenites: Magma ocean crystallization and shallow magma processes on Vesta” by B. E. Mandler and L. T. Elkins-Tanton

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Abstract—Mandler and Elkins-Tanton (2013) recently proposed an upgraded magma ocean model for the differentiation history of the giant asteroid 4 Vesta. They show that a combination of both equilibrium crystallization and fractional crystallization processes can reproduce the major element compositions of eucritic melts and broadly the range of mineral compositions observed in diogenites. They assert that their model accounts for all the howardites, eucrites, and diogenites (HEDs), and use it to predict the crustal thickness and the proportions of the various lithologies. Here, we show that their model fails to explain the trace element diversity of the diogenites, contrary to their claim. The diversity of the heavy REE enrichment exhibited by the orthopyroxenes in diogenites is inconsistent with crystallization of these cumulates in either shallow magma chambers replenished by melts from a magma ocean or in a magma ocean. Thus, proportions of the various HED lithologies and the crustal thickness predicted from this model are not necessarily valid.

The genesis of diogenites is a fundamental question for the understanding of the magmatic and differentiation history of 4 Vesta. Today, two opposite views are generally considered. Firstly, many workers consider that diogenites and eucrites are cogenetic and formed during the cooling of a global magma ocean (e.g., Ikeda and Takeda 1984; Righter and Drake 1997; Ruzicka et al. 1997; Warren 1997). Alternatively, diogenites crystallized from melts unrelated to eucrites that cooled in separate plutons, possibly in the eucritic crust (e.g., Stolper 1977; Mittlefehldt 1994; Shearer et al. 1997; Barrat et al. 2008, 2010; Yamaguchi et al. 2011). In a recent paper, Mandler and Elkins-Tanton (2013) discussed an upgraded Vestan magma ocean model. They invoked a combination of 60–70% equilibrium crystallization of a magma ocean followed by continuous extraction of the residual melt into

shallow magma chambers, where fractional crystallization combined with magmatic recharge produced all the eucritic and diogenitic lithologies. They claimed that their model is able to explain the variability of incompatible trace element concentrations in diogenites, but did not provide any calculations to support this assertion. Only major and minor (e.g., Ti, Cr) elements were modeled by the authors “because only they dictate crystallization/melting behavior, and because their initial abundances—and their abundances in HEDs—are better constrained than trace elements.” In the following, we will show that their model can be easily evaluated with some trace elements and fails to reproduce the trace element diversity of the diogenites.

Mandler and Elkins-Tanton (2013) propose a four-stage history for the early evolution of Vesta: (1) accretion and heating: development of a magma

ocean; (2) surface processing, destruction of the primitive lid, and rapid cooling and equilibrium crystallization of the magma ocean; (3) convective lockup and melt extraction; (4) and finally, fractional crystallization in shallow magma chambers generating the various types of diogenites, and eruption of eucrites.

Such a model can be tested using incompatible trace elements, and particularly the rare earth elements (REEs). In the following, we will assume that the REEs are in chondritic proportions in Vesta. This is confirmed by the ^{142}Nd systematics (e.g., Boyet and Carlson 2005). The REE abundances in the bulk silicate Vesta are fixed using the bulk compositions considered by Mandler and Elkins-Tanton (2013), assuming chondritic Ti/REE ratios. The REE abundances in the global magma ocean are consequently close to $2 \times \text{CI}$. Because all the REEs except Eu are strongly incompatible during the crystallization of olivine, chromite, low-Ca pyroxenes, and plagioclase, their behaviors are similar during fractional or equilibrium crystallization. Thus, the REE patterns of the starting and residual melts are parallel. Following Mandler and Elkins-Tanton's proposition, a two-step model combining equilibrium crystallization and fractional crystallization generates residual melts with REE abundances similar to main group eucrites after about 80% of crystallization of the system, in agreement with their calculations made with major elements. However, this simple test is of course not sufficient to validate the whole model.

The situation is less simple for cumulates. For cumulates formed from eucritic melts (ranging in composition from anorthosites, gabbros, and possibly to pyroxenites), the REE abundances of the whole rocks can be explained by the proportions of primary pigeonite and plagioclase, and by a significant contribution of trapped melt (e.g., Treiman 1997; Barrat et al. 2000). For diogenites (including the olivine diogenites or harzburgitic diogenites), a comparison between the REE abundances of their orthopyroxenes and those produced during the cooling of the magma ocean or in the shallow magma chambers is critical. Experimental works performed for a range of melt compositions and crystallization conditions (e.g., pressure and temperature) have shown that the partition coefficients of REEs for low-Ca pyroxene vary widely, but the ratios of the partition coefficients display much more limited values, particularly for Y and the heavy REEs (Fig. 1). Although the partition coefficients of the REEs for low-Ca pyroxene would certainly have changed during the crystallization of the magma ocean, their ratios were rather uniform, as exemplified by the lunar case (see table 6 in Sun and Liang [2013]). Because the residual melts formed in the Vestan magma ocean (or in the shallow magma chambers) display

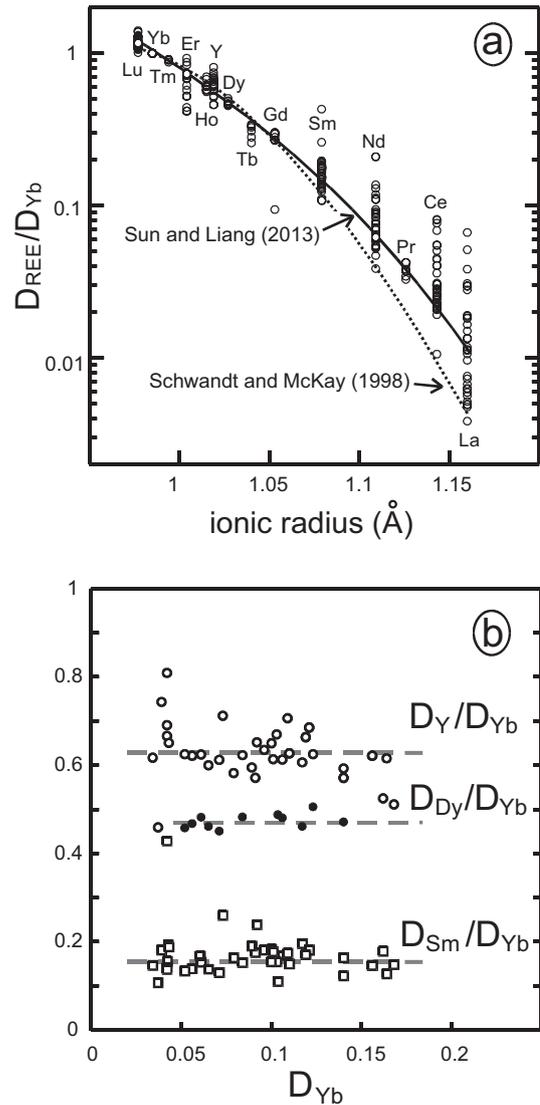


Fig. 1. Onuma diagram for ratios of experimental partition coefficients of trivalent REEs and Y for low-Ca pyroxenes (a), and selected ratios of partition coefficients for the same experiments (b). Data are from Blinova and Herd (2009), Frei et al. (2009), McDade et al. (2003a, 2003b), Salters and Longhi (1999), Salters et al. (2002), Schwandt and McKay (1998), Sun and Liang (2013), Van Kan Parker et al. (2010, 2011), and Yao et al. (2012). The scatter shown by the ratios of partition coefficients reflects, at least partly, the precision of the experimental data, which have not been filtered here. Notice that the spread of the experimental data is more limited for the heavy REE than for the light REE. Measured $D_{\text{Dy}}/D_{\text{Yb}}$ ratios close to 0.5 are consistent with parameterized lattice strain models—e.g., Schwandt and McKay (1998) and Sun and Liang (2013).

nearly chondritic REE distributions, the $(\text{Dy}/\text{Yb})_n$ ratios (normalized to CI chondrites) of the low-Ca pyroxenes are homogeneous and close to 0.5. In contrast with the homogeneity of the $(\text{Dy}/\text{Yb})_n$ ratios inferred for the magma ocean's low-Ca pyroxenes, the

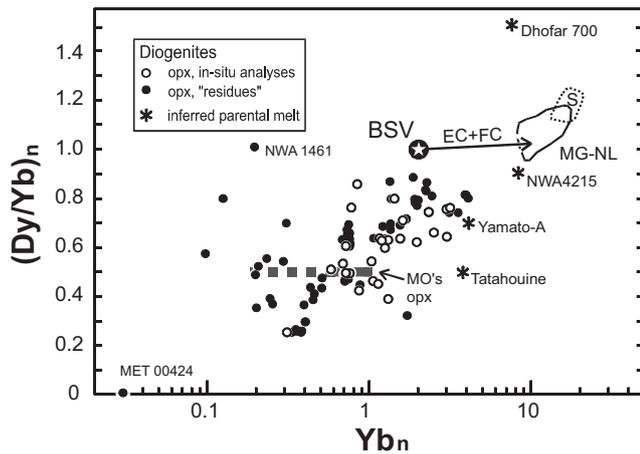


Fig. 2. $(\text{Dy}/\text{Yb})_n$ ratios versus Yb_n for eucrites—MG-NL: main group and Nuevo Laredo trend eucrites, S: Stannern trend eucrites, data from Shimizu and Masuda (1986), Barrat et al. (2000, 2007) and orthopyroxenes from diogenites (in situ analyses from Fowler et al. [1995] and Beck et al. [2013], residues after leaching from Barrat et al. [2010]). The data are normalized to CI chondrites. The trend produced by equilibrium crystallization (EC) and fractional crystallization (FC) of a melt with a starting composition equivalent to the bulk silicate Vesta (BSV) is shown for comparison. Selected partition coefficients for low-Ca pyroxene ($D_{\text{Yb}} = 0.1$ and $D_{\text{Dy}}/D_{\text{Yb}} = 0.5$) are used to calculate the compositions of the orthopyroxenes in equilibrium with the melts from the crystallizing magma ocean and from replenished shallow magma chambers (MO's opx). The same partition coefficients are used to estimate the parental melts of some diogenites—Tatahouine, Yamato-74013, NWA 4215, Dhofar 700. The pyroxenes from the diogenites and olivine (or harzburgitic) diogenites display the same range of Dy/Yb ratios, and have not been distinguished here.

range displayed by diogenitic pyroxenes is more than striking (Fig. 2). The $(\text{Dy}/\text{Yb})_n$ ratios range from 0.008 to 1.01 (e.g., Fowler et al. 1995; Barrat et al. 2010; Beck et al. 2013). MET 00424, the sample displaying the lowest value is unusual. Its composition is not understood at present. NWA 1461, the sample displaying the highest value is slightly weathered. If we omit these two extreme points, the range of $(\text{Dy}/\text{Yb})_n$ ratios is still impressive, from 0.25 to 0.89 with no gap.

The upgraded magma ocean model proposed by Mandler and Elkins-Tanton (2013) cannot explain the range of heavy REE ratios displayed by the diogenitic orthopyroxenes. Magma chamber processes are unable to explain the diversity of these ratios. For example, melts in shallow magma chambers, and those ascending from the underlying mush, display obviously parallel REE patterns. Recharging a magma chamber with these melts, as invoked by Mandler and Elkins-Tanton (2013), has no effect on the heavy-REE enrichments, and thus on the shape of the REE patterns of the

crystallizing low-Ca pyroxenes. Crustal contamination is another possible process. A mass-to-mass contamination of magma by crustal wallrocks cannot drastically modify the heavy-REE enrichment of the parental melts of eucrites because the contaminants are either poorer in REEs than the melts (contamination by cumulate eucrites or by other types of cumulates) or with a flat REE pattern (contamination by ordinary eucrites). If the wallrocks are partially melted, the resulting melts can be easily assimilated in the magma chambers. Such partial melts are rich in incompatible trace elements. Their assimilation can potentially explain the specific features of the Stannern trend eucrites (Barrat et al. 2007; Yamaguchi et al. 2009, 2013), but the increase of the Dy/Yb ratios in the contaminated eucrites was rather limited (Fig. 2). Alternatively, postcumulus processes may have been responsible for the spread of the incompatible trace elements in diogenites. These effects have been extensively discussed elsewhere (Barrat 2004; Mittlefehldt et al. 2012). The involvement of a trapped melt component could, in principle, explain an increase of the Dy/Yb ratios. However, this hypothesis fails to explain the full diogenitic range because diogenites crystallized with very low amounts of trapped melts, and because $(\text{Dy}/\text{Yb})_n$ ratio lower than 0.5 cannot be generated with this process. Subsidiary equilibration between orthopyroxenes and minor phases (e.g., plagioclase and phosphate) could have at best marginal effects on the Dy/Yb ratios of orthopyroxene, because this phase is by far the main heavy-REE carrier in diogenites. Thus, these processes cannot account for the chemical diversity of the diogenites.

At present, the most likely explanation for the observed range is a diversity of the parental melts generated on Vesta. The $(\text{Dy}/\text{Yb})_n$ ratios lower than 0.5 displayed by many diogenites, such as Tatahouine, Shalka, or Yamato-74013, point to parental melts displaying heavy REE enrichments relative to chondrites (i.e., $(\text{Dy}/\text{Yb})_n < 1$). Such melts cannot be cogenetic with eucrites. We concur with Mandler and Elkins-Tanton (2013) that the melting of the cumulates formed during the cooling of the magma ocean cannot be the parental melts of both diogenites and eucrites (see fig. 2 in Mandler and Elkins-Tanton 2013), but we reiterate that the melting of some of the magma ocean cumulates is, at present, the most likely explanation for the parental melts of the diogenites (e.g., Barrat 2004; Barrat et al. 2008, 2010; Yamaguchi et al. 2011). Finally, because a magma ocean model cannot account for the trace element features of the diogenites, it cannot be used to predict the relative proportions of the different HED lithologies. Consequently, the proportions of diogenites in Vesta and the crustal thickness estimated by Mandler and Elkins-Tanton

(2013) may not be valid, and these results should be treated with caution.

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