

IRON ISOTOPIC COMPOSITIONS OF ANGRITES AND STANNERN-TREND EUCRITES. Kun Wang¹, Frédéric Moynier¹, Nicolas Dauphas², Jean-Alix Barrat³, Paul Craddock² and Corliss Sio², ¹Department of Earth and Planetary Sciences and McDonnell Center for the Space Sciences, Washington University in St Louis, MO 63130 (wangkun@wustl.edu), ²Origins Laboratory, Department of the Geophysical Sciences and Enrico Fermi Institute, The University of Chicago, Chicago IL, ³Université de Brest, CNRS UMR 6538, I.U.E.M., France

Introduction: Recently, the development of high-resolution MC-ICP-MS has allowed measurements of iron isotopic compositions as precise as 0.03 ‰ in $\delta^{56}\text{Fe}$ [1]. Following this improvement, small yet resolvable iron isotopic variations in igneous rocks of different planetary bodies have been reported [2-8]: Terrestrial basalts are enriched in heavy iron isotopes by $\sim +0.1$ ‰ when compared to chondrites, martian and 4-Vesta (HED) meteorites, which all have similar iron isotopic compositions.

Three mechanisms have been proposed to explain this difference: (a) the bulk Earth is enriched in heavy iron isotopes following the Giant Impact that formed the Moon [2]; (b) the bulk Earth has the same iron isotopic composition as chondrites and other meteorites. However, at high-pressures (>100 GPa) relevant to terrestrial core-mantle boundary conditions, iron isotopes fractionated between metallic core and silicate mantle [9]; (c) the Fe isotopic composition of the Earth is similar to meteorites and there was no significant fractionation during core/mantle segregation. Fe(III)-bearing phases tend to be enriched in the heavy iso-

topes of iron compared to Fe(II)-bearing phases [10-13]. Therefore, terrestrial basalts, which formed from a relatively oxidized source, would have been enriched heavy isotopes of Fe [3, 14].

Achondrites are samples from differentiated planetary bodies formed under a variety of conditions and would be a good test to understand these three hypotheses for the interplanetary Fe isotopic variations. Here we report high precision isotopic compositions of Fe in angrites, martian and HED meteorites.

Samples and Method: Eight angrites, eleven martian meteorites (eight shergottites, two nakhlites and ALH84001) and thirty-two HED meteorites (twenty eucrites, three howardites and nine diogenites) were dissolved in HF/HNO₃ and Fe was purified using anion-exchange chromatography following the procedure described in [1,15]. The isotopic measurements were performed by standard-sample-bracketing on Thermo Scientific Neptune MC-ICP-MS at the University of Chicago and Washington University in St. Louis. All the data here are reported in delta notation relative to isotopic reference IRMM-014, $\delta^{56}\text{Fe} = [({}^{56}\text{Fe}/{}^{54}\text{Fe})_{\text{sample}}/({}^{56}\text{Fe}/{}^{54}\text{Fe})_{\text{IRMM-014}} - 1] \times 10^3$.

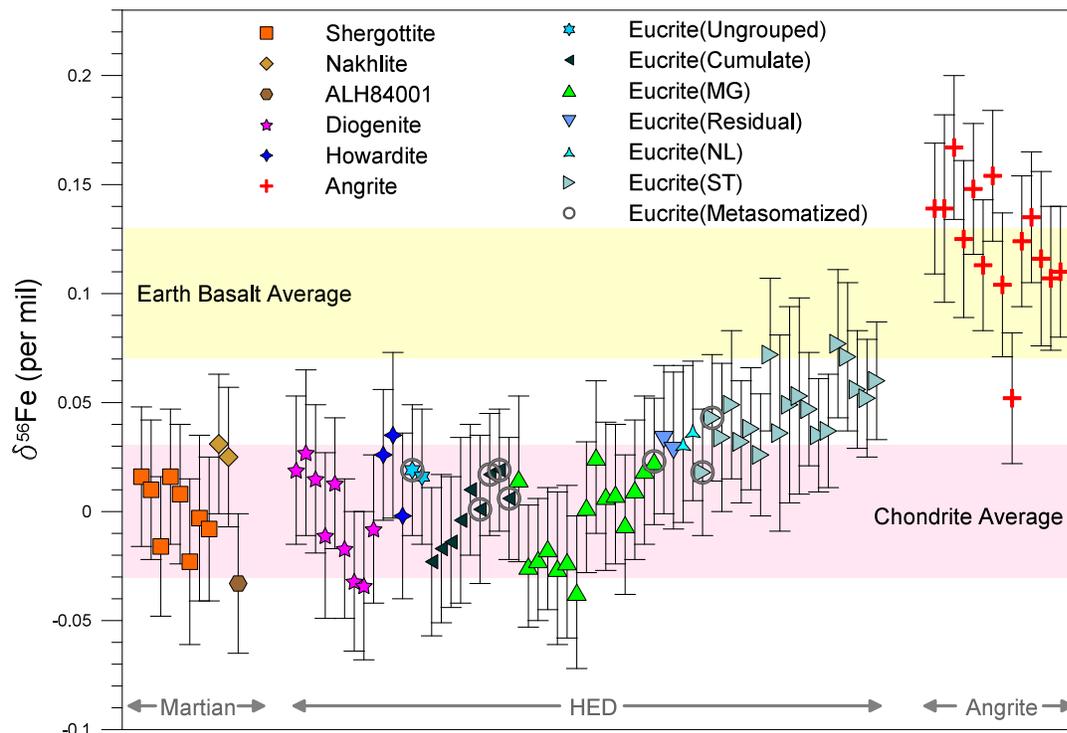


Fig. 1 Iron isotope compositions ($\delta^{56}\text{Fe}$) of martian meteorites, HEDs, and angrites. Colored shadow areas schematically show “Earth Basalt Average” and “Chondrite Average,” Eucrite (MG)=Eucrite Main Group, Eucrite (NL)=Eucrite Nuevo Laredo Trend, Eucrite (ST)=Eucrite Stannern Trend.

Results and Discussion: As shown in Fig.1, eleven martian meteorites have uniform $\delta^{56}\text{Fe}$ equal to 0.00 ± 0.01 ‰, in agreement with previous reports [2-3, 16]. The result confirms that martian meteorites have an iron isotopic composition identical to chondrites. For subgroups of HED meteorites, most eucrites (cumulate eucrites, main group, Nuevo Laredo trend and residual eucrites), all three howardites and nine diogenites also have a chondritic iron isotopic composition (weighted average $\delta^{56}\text{Fe} = 0.00 \pm 0.01$ ‰). However, it is significant that Stannern trend (ST) eucrites (weighted average $\delta^{56}\text{Fe} = +0.05 \pm 0.01$ ‰) are isotopically heavier than other HEDs. The eight samples of angrites (weighted average $\delta^{56}\text{Fe} = +0.12 \pm 0.01$ ‰) are isotopically heavier than chondrites, but similar to terrestrial basalts (see Fig.1).

Angrite parent-body is a small-sized asteroid, so it is not a viable explanation that iron isotopes could fractionate between metallic core and silicate mantle at its low pressure [9]. Since angrites are among the most oxidized meteorites (IW 1-2) [17-19], redox-controlled iron isotopic fractionation through partial mantle melting provides the best available explanation for the fractionated iron isotopic compositions of angrites. However, at IW 1-2, little Fe isotopic fractionation would be expected during partial melting, according to the quantitative model of iron isotopic fractionation between source and melt as a function of Fe(III)/Fe(II) ratio [7]. It remains to be quantified how redox conditions can influence equilibrium Fe isotopic fractionation between melts and minerals even when little ferric iron is present.

HED meteorites were formed in a low oxygen fugacity environment, close to the IW buffer [20-21]. The chondritic $\delta^{56}\text{Fe}$ values of diogenites, howardites, cumulate eucrites, main-group eucrites, Nuevo Laredo trend eucrites, and residual eucrites reflect that the $\delta^{56}\text{Fe}$ of HED parent-body mantle has not been modified by core segregation or partial melting, as is also interpreted for Mars.

The $\delta^{56}\text{Fe}$ values of ST eucrites are higher than those of other eucrites and other HED meteorites. Similar iron isotopic fractionations have been observed in lunar basalts which are also formed under reducing conditions [2-3, 22]. For lunar samples, it has been proposed that ilmenite fractionation controls the 0.1 ‰ difference in $\delta^{56}\text{Fe}$ between low-Ti and high-Ti lunar mare basalts [23]. Ilmenite are enriched in the heavy isotopes of iron by up to 0.42 ‰ in $\delta^{56}\text{Fe}$ [23]. ST eucrites are richer in Ti than other eucrites and petrography studies have also shown there are relatively more ilmenite in ST eucrites than in other eucrites [24-25]. The iron isotopic compositions of ST eucrite can

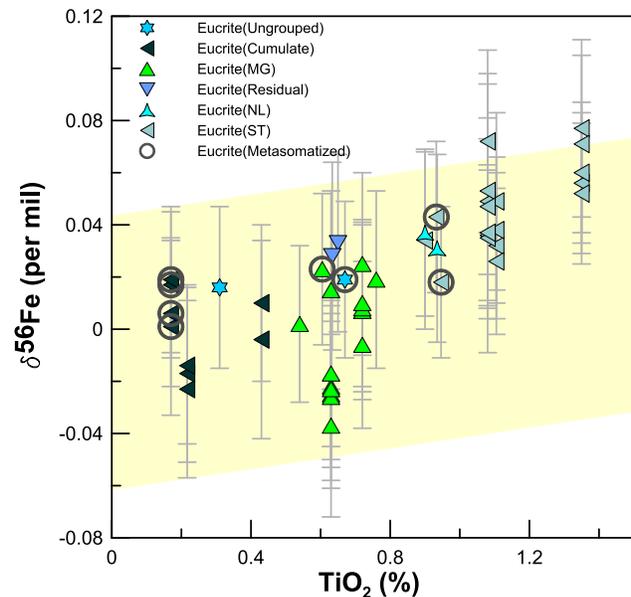


Fig. 2 Iron isotope compositions ($\delta^{56}\text{Fe}$) vs. TiO_2 in eucrites. The yellow shaded area represents ilmenite-controlled iron isotopic fractionation calculated by simple adding ilmenites. Eucrite (MG)=Eucrite Main Group, Eucrite (NL)=Eucrite Nuevo Laredo Trend, Eucrite (ST)=Eucrite Stannern Trend.

be reproduced by mixing ilmenite with $\delta^{56}\text{Fe} = +0.42$ ‰ to main-group eucrites (see yellow shaded area in Fig.2). It has been proposed recently that ST eucrites are derived from the parent magma of main-group eucrites by the assimilation of melts formed from the asteroidal crust [26]. Non-modal re-melting of previously formed asteroidal crust would preferentially consume ilmenite [26]. Therefore, partial melts formed from the crust will be richer not only in Ti, but also in the heavy isotopes of Fe. Contamination of ordinary basaltic eucrites by such components could explain the origin of ST eucrites with fractionated iron isotopic compositions.

References: [1] Dauphas, N. et al. (2009) *Chem. Geol.*, 267, 175–184. [2] Poitrasson, F. et al. (2004) *EPSL*, 223, 253-266. [3] Weyer, S. et al. (2005) *EPSL*, 240, 251–264. [4] Schoenberg, R. & von Blanckenburg, F. (2006) *EPSL*, 252, 342-359. [5] Heimann, A. et al. (2008) *GCA*, 72, 4379-4396. [6] Teng, F.Z. et al. (2008) *Science*, 320, 1620-1622. [7] Dauphas, N. et al. (2009) *EPSL*, 288, 255-267. [8] Teng, F.Z. et al. (2011) *EPSL*, 308, 317-324. [9] Polyakov, V.B. (2009) *Science*, 323, 912-914. [10] Polyakov, V.B. & Mineev, S.D. (2000) *GCA*, 64, 849-865. [11] Schauble, E.A. et al. (2001) *GCA*, 65, 2487-2497. [12] Schuessler, J.A. et al. (2007) *GCA*, 71, 417-433. [13] Shahar, A. et al. (2008) *EPSL*, 268, 330-338. [14] Williams, H.M. et al. (2004) *Science*, 304, 1656-1659. [15] Dauphas, N. et al. (2004) *Anal. Chem.*, 76, 5855-5863. [16] Anand, M. et al. (2006) *Phil. Trans. R. Soc. B*, 361, 1715-1720. [17] Jurewicz, A.J.G. et al. (1991) *Science*, 252, 695-698. [18] Jurewicz, A.J.G. et al. (1993) *GCA*, 57, 2123-2139. [19] McKay, G. et al. (1994) *GCA*, 58, 2911-2919. [20] Stolper, E. (1977) *GCA*, 41, 587-611. [21] Hewins, R.H. & Ulmer, G.C. (1984) *GCA*, 48, 1555-1560. [22] Liu, Y. et al. (2010) *GCA*, 74, 6249-6262. [23] Craddock, P.R. et al. (2010) *LPSC* 41, Abstract #1230. [24] Delaney, J.S. et al. (1984) *JGR* 89, C251-C288. [25] Warren, P.H. et al. (1990) *LPSC* 20, 281-297. [26] Barret, J.A. et al. (2007) *GCA* 71, 4108-4124.