

CAN CORE FORMATION IN PLANETESIMALS FRACTIONATE IRON ISOTOPES? CLUES FROM A STUDY OF METAL-SILICATE ASSEMBLAGES IN DISKO BASALT, GREENLAND. K. Sio¹, N. Dau-phas¹, and M. Roskosz², ¹Origins Laboratory, Department of the Geophysical Sciences and Enrico Fermi Institute, The University of Chicago, USA (ksio@uchicago.edu), ²Université des Sciences et Technologies de Lille, France.

Introduction: Magmatic iron meteorites are thought to represent core fragments from differentiated planetesimals. They possess iron isotopic compositions that are slightly heavier than chondrites [1-3], which may reflect isotopic fractionation at high temperature between metal and silicate. In order to use iron isotope as a proxy to study processes that occurred during core-mantle differentiation, it is important to know the equilibrium fractionation factor between iron metal and silicate. Polyakov and Mineev (2000) [4] presented theoretical calculations for equilibrium iron isotope fractionation between various mineral phases. Their predictions for fractionation between iron metal and Fe²⁺ bearing silicates have not yet been compared with natural systems. Pallasites may be used to study metal-silicate equilibrium fractionation [1,5-6] but several aspects of their petrogenesis are not yet well understood [7] and the iron isotope studies published so far are not well constrained. Iron metal-bearing basalts from Disko Island provide a terrestrial analogue for differentiated bodies. Their metal-silicate assemblage was the result of a reducing condition produced by the eruption of basalt through carbonaceous sediments [8]. Here, we report an iron isotope study of basalts from Disko Island.

Samples: Two samples from Disko Island, Greenland, were provided by Smithsonian Institution. Samples are fresh with no weathering.

Sample 108308-2 contains mm-sized iron blebs and its silicate portion contains ~3 wt% Fe. The large iron metal blebs could be aggregates of downward settling, smaller iron blebs [9]. Plagioclase (some skeletal), small amounts of ilmenite (some skeletal), and glass are also present in this sample.

Sample 100215 contains μ m- to mm-sized iron blebs and contains ~1 wt% Fe in the silicates. Grain size distribution is unimodal. Pyroxene, plagioclase, glass, and small amounts of troilite are found in this sample.

Methods: Metal was separated from the silicates by first grinding the sample in alcohol in a mortar. A hand magnet dipped into alcohol containing the sample preferentially removed iron metal. The grinding and metal removal was repeated until no more particles were attracted to the magnet. SEM analysis showed that the metal portions are pure iron. Grinded, final silicate grain sizes typically range from 5 to 10 μ m, though some 20 μ m grains are present. The silicate separates from the Disko basalts contain >98 % silicate grains, with the remaining 2 % being FeS and trace amounts of sub-micron sized metals grains.

Five fractions were analyzed for each Disko basalt: (1) iron metal leachates in 6 N HCl, (2) iron metal grains, (3) bulk rock, (4) silicate separates, and (5) leached silicate residue from 1 N HCl. Column chromatography method is described by [10]. Iron isotopic compositions were measured by MC-ICPMS. Results are expressed in $\delta^{56}\text{Fe}$, where $\delta^{56}\text{Fe} = [({}^{56}\text{Fe}/{}^{54}\text{Fe})_{\text{sample}} / ({}^{56}\text{Fe}/{}^{54}\text{Fe})_{\text{IRMM014}} - 1] \times 10^3$. The Mg/Fe ratio for each portion was also measured with MC-ICPMS. Measured atomic Mg/Fe ratio for geostandards RGM-2 and DRN are 0.290 and 0.886, compared to the published values of 0.285 and 0.898, respectively. A plot of $\delta^{56}\text{Fe}$ against Mg/Fe defines a mixing line for each of the Disko basalts.

To account for any iron metal still present in the silicate fraction, SEM profile analyzes were performed on the whole rock to obtain the Mg/Fe ratio for the pure silicate. For each sample, approximately 1,000 points were collected. Only those that resemble the stoichiometry of ~30 manually analyzed silicate grains and glass were taken to represent the silicate portion of that rock. The final $\Delta^{56}\text{Fe}_{\text{metal-silicate}}$ can be read from the intercepts of the mixing line with Mg/Fe = 0 and Mg/Fe = bulk silicate. Leached silicates are not used to define the mixing lines as they are more Fe-poor than bulk silicates.

Results and Discussion: The two Disko basalts show positive $\Delta^{56}\text{Fe}_{\text{metal-silicate}}$. For sample 108308-2, the fractionation is ~0.17‰, whereas in 100215, the fractionation is ~0.41‰. Closure temperatures are considered to record the temperature at which isotopes cease to exchange between phases [11]. Thus, they are used in comparing our fractionation results to the predictions of [4]. In calculating the closure temperatures, the presence of skeletal plagioclase constrains the cooling rate to approximately 2 °C/hr [12] and diffusion data of iron in diopside [13] was used. Closure temperatures for sample 100215 and 108308-2 are around 680 °C and 750 °C, respectively. These temperatures are only estimates because pyroxene grain sizes vary in the samples.

It is important to discuss whether our measured fractionations represent kinetic or equilibrium processes. Thermodynamics calculations combined with fractionation predictions of [4] allow us to predict the mixing lines for the metal-silicate system at varying temperatures (Fig. 1). Activity coefficient for FeO in melt is taken to be constant at 1.86 [14]. In this figure, Mg/Fe = 0 is the metal component of the metal-silicate assemblage, and the end point of the mixing line is the silicate component of the same assemblage.

During cooling at CCO buffer, iron can become oxidized and reenter the silicate. The length scale of iron diffusion in diopside is $l = 2(Dt)^{1/2} \sim 150 \mu\text{m}$, which is significant compared to the distance between metal grains. If kinetic processes controlled the system, the silicates are expected to incorporate lighter iron isotopes below its equilibrium value with iron metal due to faster diffusion of the lighter isotopes [15].

Sample 108308-2 defines a mixing line (Fig. 1) similar to that predicted for equilibrium at a temperature of 735 °C (Fig. 2), very close to its estimated closure temperature. However, sample 100215 shows larger $\delta^{56}\text{Fe}$ fractionation between metal and silicate than that found in 108308-2 and the reason for this difference is unknown. More Disko samples will be studied to evaluate the processes (equilibrium vs kinetic) that controlled iron isotopic fractionation between metal and silicate in Disko basalts.

Our results show that core-mantle differentiation in planetesimals may fractionate iron isotopes in planetesimals, which could explain why iron meteorites have heavy iron isotopic composition relative to chondrites. Whether such fractionation also affected larger planetary objects like the Earth is still uncertain [16-18]. The mantles of Mars and Vesta have near-chondritic iron isotopic compositions [19-20] and in those objects core formation did not fractionate measurably iron isotopic ratios.

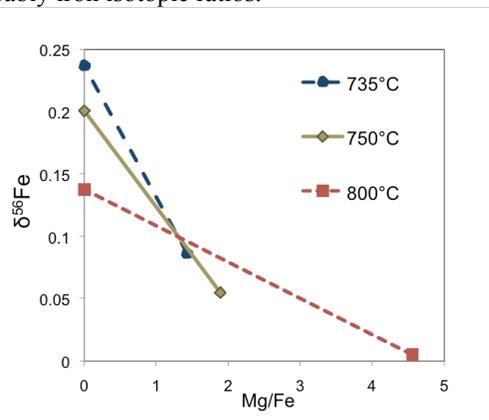


Fig. 1. Calculated mixing lines. Atomic Mg/Fe ratios for the silicates are obtained from thermodynamic calculations at CCO buffer. $\delta^{56}\text{Fe}$ values are from Polyakov and Mineev (2000) [4]. Bulk composition is chosen to be at $\delta^{56}\text{Fe} = 0.1 \text{‰}$, and bulk atomic Mg/Fe = 1.3.

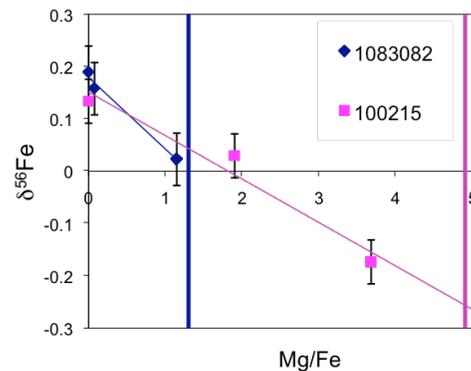


Fig. 2. Mixing lines for each Disko basalt defined by iron, bulk, and silicate portions of the Disko basalts. The vertical lines represent the bulk Mg/Fe ratios of the pure silicates measured by SEM.

References: [1] Poitrasson F. et al. (2005) *Earth Planet Sc. Lett.* 234, 151-164. [2] Dauphas N. et al. (2009) *Earth Planet Sc. Lett.*, in press [3] Williams H.A. et al. (2006) *Earth Planet Sc. Lett.*, 250, 486-500. [4] Polyakov V.B. & Mineev S.D. (2000) *Geochim. Cosmochim. Ac.* 64, 849-865. [5] Zhu, X.K. et al. (2002) *Earth Planet Sc. Lett.*, 200, 47-62. [6] Weyer S. et al. (2005) *Earth Planet Sc. Lett.*, 240, 251-264. [7] Haack, H. (2003) *Treatise on geochemistry* 1, p.325-345, Elsevier. [8] Goodrich C.A. & Bird J.M. (1985) *J. Geol.* 93, 475-492. [9] Klock W. et al. (1986) *Contrib. Mineral Petr.* 93, 273-282. [10] Dauphas, N. et al. (2004) *Anal. Chem.* 76, 5855-5863. [11] Dodson, M.H. (1973) *Contrib. Mineral Petr.* 40, 259-274. [12] Lofgren, G. (1974) *Am. J. Sci.* 274, 243-273. [13] Azough R. & Freer R. (2000) *Phys. Chem. Miner.* 27, 732-740. [14] Ottonello G. et al. (2001) *Chem. Geol.* 174, 157-179 [15] Richter F.M. et al. (2009) *Chem. Geol.* 258, 92-103 [16] Poitrasson F. et al. (2009) *Earth Planet Sc. Lett.* 278, 376-385 [17] Polyakov V.B. (2009) *Science* 323, 912-914 [18] Shahar, A. et al. (2009) *LPS 40, Abstract #1640*. [19] Poitrasson F. et al. (2004) *Earth Planet Sc. Lett.* 253-266 [20] Anand, M et al. (2006) *Philos. T. R. Soc. B* 361, 1715-1720.