

EQUILIBRIUM METAL-SILICATE Fe ISOTOPE FRACTIONATION AND THE IMPLICATIONS FOR PLANETARY DIFFERENTIATION M. K. Jordan¹ and E. D. Young¹, ¹Department of Earth, Planetary, and Space Sciences, University of California, Los Angeles, USA, mkjordan@ucla.edu, eyoung@epss.ucla.edu.

Introduction: Variations in Fe isotope composition are observed among different planetary bodies. The Earth and Moon are isotopically heavy in $^{57}\text{Fe}/^{54}\text{Fe}$ relative to chondrites, HED meteorites (Vesta), and Mars [1]. Several mechanisms have been suggested to try and account for these variations. Possible explanations include evaporation after the Moon-forming impact, igneous differentiation, or core formation. Magmatic iron meteorites exhibit an enrichment in $^{57}\text{Fe}/^{54}\text{Fe}$ relative to chondrites of $\sim 0.2\%$ [2]. Though not definitive, this is suggestive of heavy Fe partitioning into the cores of differentiated bodies. We aim to determine if core formation is accompanied by an isotopic signature. Understanding equilibrium Fe isotope fractionation between metal and silicate phases is fundamental to assessing the significance of the variation of Fe isotopes in the Solar System.

Previous work has shown that there is a measurable Si isotope fractionation between silicate and metal [3]. This effect was verified by comparing fractionation in the laboratory with that in enstatite achondrites [4]. One might expect to observe a similar effect for Fe isotopes due to fractionation between metal and silicate.

Measurements of Fe isotope fractionations between phases in meteorites, and measurements in the laboratory, have yielded contradictory results for $^{57}\text{Fe}/^{54}\text{Fe}$ equilibrium fractionation between metal and silicate. Using the metal and silicate phases of pallasites, [2] and [5] found the metal to be high in $^{57}\text{Fe}/^{54}\text{Fe}$ compared to the olivine grains, while [6] found no systematic difference between the two phases (Figure 1). The equilibration of Fe isotopes between metal and olivine in pallasites is debated [2, 6]. Experimental results have failed to provide a more definitive answer. Shahar et al. [7, 8] used piston cylinder experiments and found the metal to be isotopically heavy relative to silicate, while [2] and [9] found no significant fractionation, even at low temperatures.

Here we determine the equilibrium Fe isotope fractionation between the metal and silicate phases using aubrites Norton County and Mount Egerton.

Sample Description: Both Norton County and Mount Egerton are enstatite achondrites (aubrites). Norton County is a brecciated aubrite, while Mount Egerton is an unbrecciated anomalous aubrite. Both rocks consist primarily of nearly FeO-free enstatite with forsterite, diopside, and plagioclase comprising the remaining silicate material. The low Fe content in

the silicate material is a result of the reducing conditions under which aubrites formed and renders analysis of the silicate phase difficult. The metal is primarily kamacite, with minor amounts of perryite, shreibersite, and troilite present. In Norton County, the metal occurs mainly as irregular shaped grains and accounts for about 1-1.5 vol% of the meteorite [10, 11]. In Mount Egerton, the metal accounts for 21 wt% and occurs as cm-sized blobs [10].

Ziegler et al. [4] determined the $^{29}\text{Si}/^{28}\text{Si}$ fractionation between metal and silicate phases for Norton County and Mount Egerton. The agreement between the Si isotope temperatures calibrated in the laboratory and previous Si-concentration thermometry suggests that these rocks are in isotopic equilibrium at a temperature of 1130 K and $1200 \text{ K} \pm 80 \text{ K}$, respectively [3, 4, 12].

Uncertainties in activity coefficients and oxygen fugacities limit the accuracy of the equilibration temperatures obtained from Si concentrations in metal [4, 12]. We calculate an $f\text{O}_2$ for these rocks of $\sim \text{IW} - 5.2$ based on the Fe concentration in enstatite. An independent check on the temperature using this oxygen fugacity requires an estimate for the activity coefficient of FeO in the enstatite-rich melts. This work is ongoing

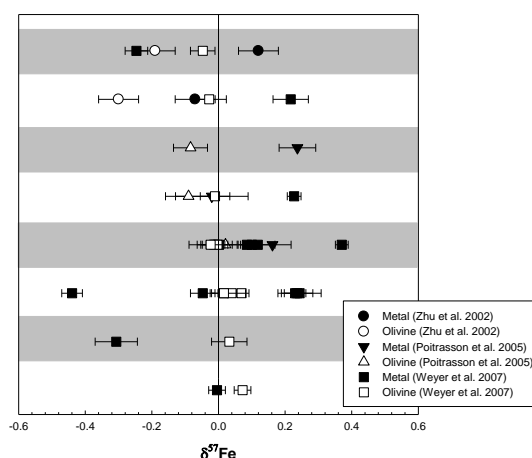


Figure 1. Compilation plot of Fe isotope compositions for metal and olivine separates from Pallasites.

Analytical Methods: Large grains of enstatite were crushed and metallic grains and sulfides subsequently removed. Silicates were dissolved on a hotplate in a mixture of concentrated HF and HNO_3 . Metals were dissolved in concentrated HCl.

Ion-exchange chromatography was used to separate Fe. Metal samples were loaded on 0.3 mL of X8 resin (100-200 mesh, chloride form). Matrix elements were eluted using 8 N HCl and Fe was eluted using 0.5 N HCl. At minimum, we require ~10 μg of Fe to make precise measurements. The Mg/Fe ratio is 500:1 in the silicate material from Norton County, requiring ~20 mg of silicate to obtain sufficient Fe from the silicate. As a consequence, a different column procedure was used to process the large amount of material for the silicate portion. Fe-poor silicate samples were loaded onto columns consisting of 1 mL AG 1-X8 resin (200-400 mesh, chloride form). Matrix elements are eluted using 6 N HCl and Fe is eluted using 0.4 N HCl.

Data were collected on a ThermoFinnigan NeptuneTM multiple-collector inductively coupled plasma-source mass spectrometer (MC-ICP-MS) run in wet plasma mode. Samples were run at a mass resolving power (instrumental $\Delta m/m$) of >9000 to resolve ArO^+ among other isobaric interferences. Corrections for instrumental mass bias were made using sample-standard bracketing and peak height matching between samples and standards. Samples were measured against our SPEX CertiPrep[®] 2 standard, which is referenced to IRMM-14.

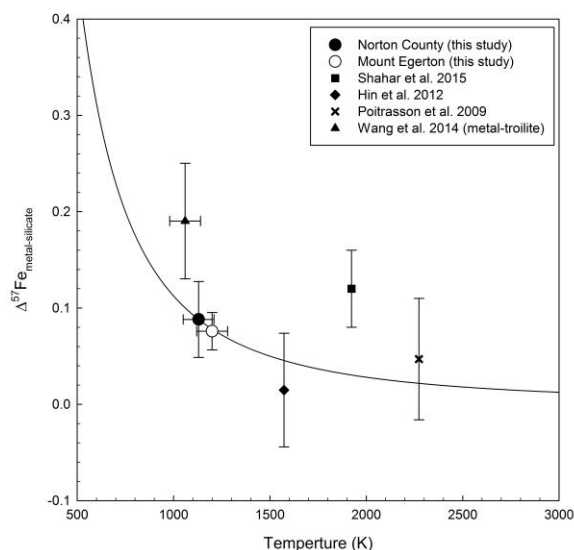


Figure 2: Temperature calibration determined from measured metal-silicate fractionation of Norton County and Mount Egerton. Experimental and natural sample results for metal-silicate fractionation plotted.

Results: Data are reported as per mil deviations from the standard IRMM-14. The average metal $\delta^{57}\text{Fe}$ values for Norton County and Mount Egerton are $0.030\text{‰} \pm 0.035$ (2 SE) and $0.024\text{‰} \pm 0.015$ (2 SE). The average silicate $\delta^{57}\text{Fe}$ values are $-0.058\text{‰} \pm 0.019$ (2 SE) and $-0.052\text{‰} \pm 0.012$ (2 SE), respectively. The

calculated isotopic fractionation $\Delta^{57}\text{Fe}_{\text{metal-silicate}}$ is $0.08\text{‰} \pm 0.039$ (2 SE) for Norton County and $0.09\text{‰} \pm 0.019$ (2 SE) for Mount Egerton, indicating that the heavy isotopes of Fe partition into the metallic phase.

We can use the equilibration temperatures for these rocks in conjunction with the equilibrium Fe isotope fractionation between metal and silicate to establish a temperature calibration since isotope fractionation varies as $1/T^2$ (Figure 2). Experimental results are plotted for comparison as well as the fractionation between metal and troilite measured by Wang et al. [13] Metal-silicate and metal-troilite fractionation should be very similar [14, 15]

Discussion: Using our calibration, one calculates the $\delta^{57}\text{Fe}$ of the BSE to be approximately 0.002‰ below chondritic $\delta^{57}\text{Fe}$. This value is below our current level of detection. As a result, equilibrium Fe fractionation during core formation is probably not responsible for the heavy Fe isotope compositions observed in the Moon and Earth relative to chondrites. The calibration should prove more useful for smaller bodies that formed under lower pressure and temperature conditions. One potential application is Vesta, which formed a metallic core at lower pressures and temperatures less than 2000 K [16]. Assuming a temperature of ~1800 K for Vestan core formation, we predict a metal-silicate fractionation of about 0.035‰ , with the bulk silicate having a $\delta^{57}\text{Fe} \sim -0.024\text{‰}$ relative to bulk Vesta (e.g., chondrite). This number is at our current level of precision and still difficult to measure. Our results thus far suggest that the 0.2‰ higher $^{57}\text{Fe}/^{54}\text{Fe}$ of iron meteorites cannot be explained as the result of silicate-metal partitioning of chondritic like bodies.

References: [1] Poitrasson F. et al. (2004) *EPSL*, 223, 253-266. [2] Poitrasson, F. et al. (2005) *EPSL*, 234, 151-164. [3] Shahar A. et al. (2011) *GCA*, 75, 7688-7697. [4] Ziegler K. et al. (2010) *EPSL*, 295, 487-496. [5] Zhu X. K. et al. (2002) *EPSL*, 200, 47-62. [6] Weyer S. et al. (2005) *EPSL*, 240, 251-265. [7] Shahar et al (2012) *LPS XLIII*, Abstract #2049. [8] Shahar et al. (2013) *LPS XLIV*, Abstract #2351. [9] Hin R. C. et al. (2012) *GCA*, 77, 164-181. [10] Watters T. R. and Prinz M. (1979) *LPS X*, 1073-1093. [11] Casanova I. et al. (1993) *GCA*, 57, 675-682. [12] Wasson J. T. et al. (1994) *Meteoritics*, 29, 658-662. [13] Wang K. et al. (2014) *GCA*, 142, 149-165. [14] Polyakov V.B. et al. (2007) *GCA*, 71, 3833-3846. [15] Polyakov, V.B. and Mineev S.D. (2000) *GCA*, 64, 849-865. [16] Zhou Q. et al. (2013) *GCA*, 110, 152-175.