

IRON ISOTOPE CONSTRAINTS ON PLANETESIMAL CORE FORMATION IN THE EARLY SOLAR SYSTEM. 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: The prevalence of metallic planetary cores in the Solar System indicates that core formation is a fundamental process in planet building, understanding core-mantle differentiation is therefore crucial to understanding planetary formation and evolution.

The Fe isotope compositions of Earth, Moon, Mars (SNC meteorites), Vesta (HED meteorites), and several meteorite classes have been previously measured and exhibit considerable variability [1-10]. One possible explanation for this variability is differences in the processes that produced the cores of these bodies. Alternative explanations include evaporation and igneous differentiation.

Magmatic iron meteorites are thought to represent the cores of differentiated asteroid-sized bodies. The meteorite record indicates that these metallic cores have high $^{57}\text{Fe}/^{54}\text{Fe}$ with respect to chondritic composition by $\sim 0.12\%$ [1-9]. This difference between chondrites and magmatic iron meteorites is potentially explained if the heavy isotopes of Fe partition into the metallic melt during planetary differentiation and core formation. While the observation indicates that this may be the case, uncertainty arises due to the probable lack of a cogenetic origin for these rocks.

To explore the likelihood of the core-formation mechanism as the cause of the observed trends, we measured the iron-isotope compositions of the metal and silicate phases in two aubrite meteorites to determine metal-silicate equilibrium Fe isotope fractionation. Additionally, we measured the $^{57}\text{Fe}/^{54}\text{Fe}$ of several iron meteorites to determine if fractional crystallization affects Fe isotope composition. Previously, it has been shown that $^{102}\text{Ru}/^{99}\text{Ru}$ of magmatic iron meteorites increases with increasing degree of fractional crystallization [11].

Samples: We measured Fe isotope ratios in metal and silicate coexisting in two aubrite meteorites (enstatite achondrites), Norton County and Mount Egerton. Estimated temperatures of equilibration for these rocks allow us to establish a temperature calibration for equilibrium Fe-isotope fractionation between metal and silicate.

We also measured six IIIAB iron meteorites, with varying degrees of fractional crystallization evidenced in their incompatible element concentrations. Grant, Mount Edith, and Buenaventura experienced high degrees of fractional crystallization while the less evolved Haig, Henbury, and Kenton County experienced rela-

tively low degrees of fractional crystallization. Grant and Henbury are among those meteorites analyzed in the Ru isotope study [11].

Analytical Methods: Ion-exchange chromatography was used to purify Fe. Metal samples were loaded on 0.3 mL of AG1-X8 resin (100-200 mesh, chloride form). Matrix elements were eluted using 9 N HCl and Fe was eluted using 0.5 N HCl. Silicate samples required a large amount of material due to the low concentration of Fe in the silicate phases. 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.

Results: Data are reported as per mil deviations from the standard IRMM-14. The average $\delta^{57}\text{Fe}$ values for the metal fractions from Norton County and Mount Egerton are $0.030\% \pm 0.035$ (2 SE) and $0.024\% \pm 0.015$ (2 SE), respectively. The average $\delta^{57}\text{Fe}$ values for the silicate fractions are $-0.058\% \pm 0.019$ (2 SE) and $-0.052\% \pm 0.012$ (2 SE), respectively. The calculated metal-silicate isotopic fractionation, $\Delta^{57}\text{Fe}_{\text{metal-silicate}}$, is $0.08\% \pm 0.039$ (2 SE) for Norton County and $0.09\% \pm 0.019$ (2 SE) for Mount Egerton. The positive value indicates that the heavy isotopes of Fe preferentially partition into the metallic phase during metal-silicate differentiation.

The $\delta^{57}\text{Fe}$ values for the IIIAB iron meteorites range from $+0.061$ to $+0.169\%$ (± 0.02 2 SE). The measured values fall within the range of values seen in the literature for both magmatic iron meteorites overall and for IIIAB iron meteorites specifically [1-9].

Discussion: The $\delta^{57}\text{Fe}$ values of the iron meteorites are plotted against the degree of fractional crystallization for those meteorites in Figure 1. We use the relative concentration of the highly incompatible element As as a measure of the degree of fractional crystalliza-

tion. Because of uncertainties in values for α , several Rayleigh fractionation curves are plotted along with the data in Figure 1. The different curves represent the case where the heavy isotopes of Fe partition into the solid metal ($\alpha = 1.0002$) and the case where the lighter isotopes of Fe partition into the solid metal phase ($\alpha = 0.9998$). The plot shows that there are no systematic trends among these magmatic iron meteorites; iron meteorites that crystallized late (Buenaventura, Grant, and Mt. Edith) span the same ranges of values for $\delta^{57}\text{Fe}$ as those that crystallized early (Henbury, Haig, and Kenton County), indicating that fractional crystallization did not affect the Fe isotope composition of the magmatic iron meteorites.

The solubility of Si in metal is temperature dependent and can be used to estimate the temperature of equilibration for the aubrite meteorites. We use activity models from [12] to estimate the temperature of equilibration, inputting parameters (f_{O_2} , γ_{Fe}) suitable for aubrite meteorites. We estimate equilibration temperatures of ~ 1415 K for Norton County and 1460 K for Mount Egerton.

Taking into account the non-Fe constituents in the metal phase, which affects equilibrium Fe stable isotope fractionation [13] and the temperature of equilibration for the aubrite meteorites we establish a temperature calibration. We use our temperature calibrations in conjunction with mass balance, to calculate the $\delta^{57}\text{Fe}$ values for the silicate and metallic phases of a body as a function of t where core and mantle equilibrated at 1800 K (i.e. Vesta) as shown in Figure 2.

The pair of lines in Figure 2 represent $\delta^{57}\text{Fe}$ values for coexisting metal and silicate as a function of the fraction of iron contained in the silicate portion of the body. The dotted lines represent the case where there is 8 wt. % Ni in the core and $\alpha_{\text{core/mantle}} = 0.17\text{‰}$. The solid lines represent the case where there is 10 wt. % Ni in the core and $\alpha_{\text{core/mantle}} = 0.20\text{‰}$. These Ni concentration values represent those typical for IIIAB iron meteorites. For a body with a core/mantle mass ratio like the Earth, the silicate should be low in $\delta^{57}\text{Fe}$ and metal should be nearly chondritic because the core dominates the iron budget. Conversely, smaller cores with greater fractions of the Fe in the silicate mantles increase the $\delta^{57}\text{Fe}$ of the core to values greater than chondritic. Even with extremely small core sizes and impurities consistent with those in iron meteorites, the observed elevated $^{57}\text{Fe}/^{54}\text{Fe}$ among magmatic iron meteorites relative to chondrite is difficult to explain using the fractionation between silicate and metal that we observe in the meteorites themselves.

Although our results indicate that the heavy nuclides of Fe preferentially partition into the metal phase, the magnitude of this fractionation cannot ac-

count for the observed variability among Solar System materials. Instead evaporation or some combination of core formation and evaporation of planetesimals may explain the variety we see in the Fe isotope compositions of Solar System materials.

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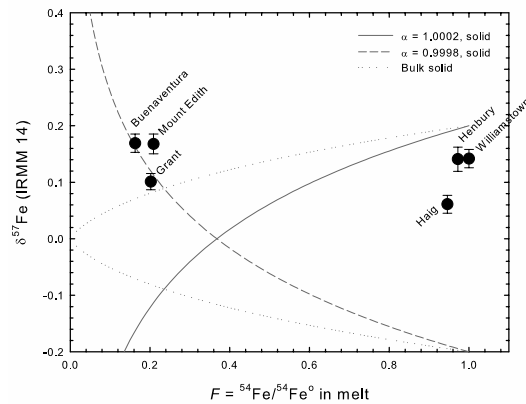


Figure 1. Fe isotope composition of IIIAB iron meteorites as a function of degree of fractional crystallization.

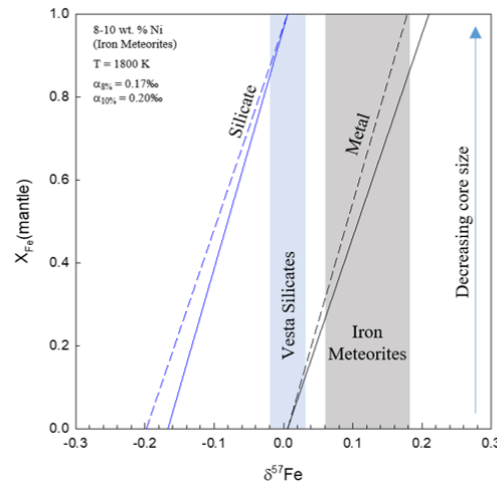


Figure 2. $\delta^{57}\text{Fe}$ of coexisting metal and silicate phases as a function of core size as a result of the mole fraction of Fe that is present in the core vs. mantle.