

IRON ISOTOPE COMPOSITION OF THE LUNAR MARE REGOLITH: IMPLICATIONS FOR ISOTOPIC FRACTIONATION DURING PRODUCTION OF SINGLE DOMAIN IRON METAL. R. A. Wiesli¹, B. L. Beard¹, L. A. Taylor², S. A. Welch¹, C. M. Johnson¹, ¹Department of Geology and Geophysics, University of Wisconsin, Madison WI 53706, ²Planetary Geosciences Institute, University of Tennessee, Knoxville TN 37996.

Introduction: The processes of space weathering are dominant in the development of the regolith and soils on all airless bodies in the Solar System. Indeed, it is this processing of the soil that complicates and masks the effective use of remote sensing by spectral reflectance [1-2]. The lunar regolith provides us with invaluable samples to study these effects of space weathering. In particular, variations in isotopic compositions can constrain some of the processes that act on the lunar soils. It has been observed that formation and maturation of the lunar regolith enrich the heavy isotopes of O, Si, S, K, Cd, Ca, and Mg, in comparison to lunar rocks [3-8], although the effect on Ca and particularly Mg are only minor. The mass-dependent isotopic variations of these elements are variably explained by: 1) ion sputtering [9]; 2) volatilization by micrometeorite impacts [3, 10]; and 3) redeposition of some sputtered or volatilized matter after gravitational fractionation [11]. The relative contributions of these processes to the isotopic fractionations are not well known. However, the abundance of nanophase Fe in lunar soils can provide signatures of some of these competitive processes.

Analytical Methodology: We report here the first Fe isotope analyses of lunar mare regolith samples. These samples are from the studies of the Lunar Soil Characterization Consortium, where a selection of mare soils were sieved into sized fractions (<10 μm , 10-20 μm , and 20-45 μm) [1]. In this present study, six of these mare soils were analyzed for their Fe isotope compositions. In addition, the <10 μm size fraction of 79221 was partially leached with 1M HCl, and the leachate (63% of total Fe), and residue were analyzed for their Fe isotope compositions. All analyses were performed using the Univ. of WI *IsoProbe* [12-13]. Based on replicate analyses of samples processed through the entire analytical procedure two or more times, the average reproducibility for the $^{56}\text{Fe}/^{54}\text{Fe}$ is $\pm 0.05\text{‰}$. Analyses are reported using delta notation, where: $\delta^{56}\text{Fe} = ([^{56}\text{Fe}/^{54}\text{Fe}]_{\text{sample}}/[^{56}\text{Fe}/^{54}\text{Fe}]_{\text{Terr Ig rocks}} - 1)10^3$ and $\delta^{57}\text{Fe} = ([^{57}\text{Fe}/^{54}\text{Fe}]_{\text{sample}}/[^{57}\text{Fe}/^{54}\text{Fe}]_{\text{Terr Ig rocks}} - 1)10^3$ [12]. On this scale the measured Fe isotope composition of the IRMM-014 Fe isotope standard was $\delta^{56}\text{Fe} = -0.09 \pm 0.05\text{‰}$ and $\delta^{57}\text{Fe} = -0.11 \pm 0.07\text{‰}$.

Results: The Fe isotope composition of the various size fractions for the different soils are plotted in Figure 1 versus the I_s/FeO values determined for each size fraction of each soil [1]. There is an overall positive correlation between the soil maturity index I_s/FeO [14] and $\delta^{56}\text{Fe}$.

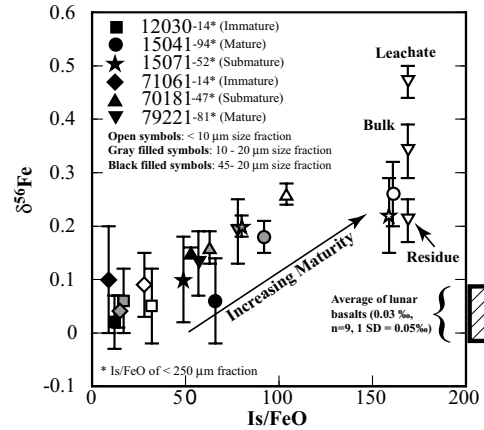


Figure 1. Plot of I_s/FeO vs. $\delta^{56}\text{Fe}$ for lunar soils. All analyses are for bulk samples of the different size fractions. In addition, the <10 μm size fraction of 79221 has been leached and the $\delta^{56}\text{Fe}$ values of the leachate and the residue are given. For comparison, the range for $\delta^{56}\text{Fe}$ for lunar basalts are shown in the lower left corner. Error bars indicate 1 SD for duplicate or triplicate analyses.

The I_s/FeO values indicate the relative percentage of the total FeO of a sample that is present as nanophase Fe metal. Furthermore, the highest values of I_s/FeO occur in the finest size fractions of lunar soil [1, 14]. Soils 71061 and 12030 (immature) show no correlation with isotopic composition. The leached aliquot of the most mature soil (<10 μm size fraction) has the highest I_s/FeO ratio, and also the highest $\delta^{56}\text{Fe}$ value ($0.47 \pm 0.03\text{‰}$). On a three-isotope plot, all analyses plot along a line with slope 2/3, indicating that the fractionations observed are associated with mass-dependent behavior (Figure 2).

Lunar Rock Analyses: In addition to these Fe isotope analyses of lunar regolith samples, we have reanalyzed the five Apollo 12 mare basalt samples reported in Beard et al. [15], as well as 3 additional mare basalts (2 from the Apollo 12 landing site and 1 from the Apollo 17 landing site), and 1 impact melt rock from the Apollo 16 landing site. The Apollo 12 mare basalts have $\delta^{56}\text{Fe}$ values that range from -0.03 to $+0.07\text{‰}$, the Apollo 17 mare basalt has a $\delta^{56}\text{Fe}$ value of 0.13‰ , and the impact melt rock has a $\delta^{56}\text{Fe}$ value of -0.04‰ . Although the Apollo 17 mare basalt yielded a $\delta^{56}\text{Fe}$ value of $0.13 \pm 0.02\text{‰}$, similar to the value reported by Poitrasson et al. [16], i.e., $0.07 \pm 0.04\text{‰}$ (data corrected through a common value for the IRMM-014 Fe isotope standard), Apollo 12 rocks and the Apollo 16 impact rock have lower $\delta^{56}\text{Fe}$ values ($0.02 \pm 0.04\text{‰}$ and $-0.04 \pm 0.02\text{‰}$, respectively).

Discussion: The iron isotope composition of lunar mare basalts (8 samples) and 1 impact melt rock have

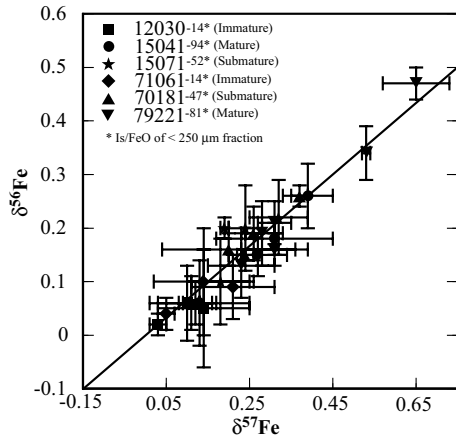


Figure 2. Plot of $\delta^{57}\text{Fe}$ vs. $\delta^{56}\text{Fe}$ for the lunar soils. Data points plot along the line with slope 2/3, indicating that isotopic variations are caused by mass-dependent processes. Error bars show 1 SD about the mean for duplicate analyses.

$\delta^{56}\text{Fe}$ values that range from -0.04 to $+0.13$ ‰. This is a range similar to that of terrestrial igneous rocks [12]. Immature soils 71061 and 12030 that have I_s/FeO values less than ~ 50 also have Fe isotope compositions that overlap with the average Fe isotope composition of lunar rocks. In contrast, more mature lunar soils that have high I_s/FeO values, have $\delta^{56}\text{Fe}$ values that are greater than lunar rocks. The positive correlation between I_s/FeO and $\delta^{56}\text{Fe}$ values is taken as strong evidence that the lunar regolith phase with high $\delta^{56}\text{Fe}$ values is nanophase Fe metal. Moreover, the 1M HCl leaching experiment performed on the < 10 μm size fraction of soil sample 79221 is consistent with this hypothesis. HCl preferentially attacks Fe metal; the leachate has a $\delta^{56}\text{Fe}$ value of $+0.47$ ‰ and contains 63 mole % of the total Fe in this sample. The residue from this leaching experiment has the same Fe isotope composition ($\delta^{56}\text{Fe} = 0.21$ ‰) as the coarser sized fractions of sample 79221. Assuming a linear interpolation and a target-rock Fe-isotope composition, the same as the average of lunar rock samples, the nanophase Fe metal would have a $\delta^{56}\text{Fe}$ value of $+0.8$ ‰.

Recent TEM studies by Keller & McKay [17-18] have shown that nanophase Fe metal is present in thin surface layers on many soil particles. It is with the finest soil particles of mature mare soils that the surface area versus mass becomes so significant that the majority of the nanophase Fe metal is on the surface of the soil particles [1]. Most of these patinas are the result of vapor deposition caused by micrometeorite impact [1], although a contribution from ion sputtering is also possible [19]. We suggest that during micrometeorite impact, vaporization of FeO occurs, accompanied by dissociation to elemental Fe. Both during this process and with subsequent condensation, Fe is isotopically fractionated. It would appear that some of the isotopically lighter Fe was preferentially lost from the moon [1, 3].

It has been pointed out that volatility of an element and chemical mass bias may play a role in the fractionation behavior during meteorite impact [5]. During meteorite impacts, the more volatile elements such as Na, K, Si, and Fe will be preferentially vaporized in comparison to the more refractory elements (i.e., Ca, Al, and Ti) [20]. TEM studies on the finest size fraction of lunar soils have shown a strong enrichment in the volatile elements Si and Fe, thus supporting deposition from impact-produced vapors [17].

Iron isotope compositions of the lunar soils may potentially be altered by extra-lunar iron derived from meteorite impacts. Zhu et al. [21] have analyzed a variety of meteorite types, obtaining a range of $\delta^{56}\text{Fe}$ of -0.52 ‰ to 0.72 ‰ relative to igneous rocks (data corrected through a common value for the IRMM-014 Fe isotope standard). From this range we conclude that it would be unlikely that the trend observed in our analyses could be caused by a meteoritic contribution, i.e., the enrichment of $\delta^{56}\text{Fe}$ in the < 10 μm fraction of mature soils is a signature of space weathering processes that are active on the lunar surface. The question remains to how this Fe isotope fractionation relates to the formation of all nanophase Fe in the soils. The axiom that the nanophase Fe in the agglutinitic glass formed as a result of impact-generated melting of solar-wind saturated lunar soils has recently been questioned by Taylor et al. [22]. It was speculated on that the majority of the nanophase Fe in the soils was formed by vapor deposition and subsequent reworking of the soil. In any case, the fractionation of the Fe isotopes was most likely the result of some Fe vaporization, possibly supporting this new paradigm [22].

References: [1] Taylor L. A. et al. (2001) *JGR*, 106, 27985-27999. [2] Pieters C. M. & Fischer E. M. (1993) *JGR*, 98, 20817-20824. [3] Clayton R. N. et al. (1974) *PLSC*, 1801-1809. [4] Thode H. G. & Rees C. E. (1976) *PLSC*, 459-468. [5] Humayun M. & Clayton R. N. (1995) *GCA*, 59, 2115-2130. [6] Sands D. G. et al. (2001) *LPSL*, 186, 103-111. [7] Russel W. A. et al. (1977) *PLSC*, 3791-3805. [8] Esat T. M. & Taylor S. R. (1992) *GCA*, 56, 1025-1031. [9] Switkowski Z. E. et al. (1977) *JGR*, 82, 3797-3804. [10] Hapke B. et al. (1975) *Moon*, 13, 339-353. [11] Housley R. M. (1979) *PLPSC*, 1673-1683. [12] Beard B. L. et al. (2003) *Chem. Geology*, in press. [13] Skulan J. L. et al. (2002) *GCA*, 66, 2995-3015. [14] Morris, R. V. (1976) *PLSC*, 315-335. [15] Beard B. L. et al. (2001) *LPSC XXXII*, abstr. # 1831. [16] Poitrasson F. et al. (2002) *Eos. Trans. AGU*, 83 (47), abstr. P11A-0355. [17] Keller L. P. & McKay D. S. (1997) *GCA*, 61, 2331-2341. [18] Keller L. P. & McKay D. S. (1993) *Science*, 261, 1305-1307. [19] Bernatowicz T. J. (1994) *LPSC XXVII*, 105-106. [20] De Maria G. et al. (1971) *PLSC*, 1367-1380. [21] Zhu X. K. et al. (2001) *Nature*, 412, 311-313. [22] Taylor L. A. et al. (2001) *The Moon Beyond 2002*, LPI, abstr.# 3022.