

MODERATELY VOLATILE ELEMENTS (MVEs) IN LUNAR BASALTS; AN INVESTIGATION INTO THE EVOLUTION OF THE MARE SOURCE REGIONS. Burney, D.^{1,2}, Neal, C.R.^{1,2}, and G. Jeffrey Taylor^{2,3}.

¹University of Notre Dame, Notre Dame IN, 46556; ²Center for Lunar Science and Exploration, Lunar and Planetary Institute, Houston, TX 77058; ³Hawaii Institute of Geophysics and Planetology, University of Hawaii, Honolulu, Hawaii 96822; dburney@nd.edu; neal.1@nd.edu

Introduction: The Giant Impact hypothesis is the most commonly accepted mechanism of Moon formation [1,2]. It posits that a Mars-sized planetesimal collided with proto-Earth, and the ejected debris (sourced from both the impactor and Earth) subsequently coalesced and cooled to form the Moon [2]. The violence of this process would have created temperatures capable of volatilizing a great deal of material, and the Moon was hypothesized to be depleted highly volatile species (eg, H₂O, CO_x, F, Cl, and S) [3]. Recent analyses on olivine-hosted melt inclusions, volcanic glass beads, and nominally hydrous mineral phases (apatite) have shown this “dry Moon” hypothesis to be incorrect [4-6]. While relative to Earth the Moon is depleted in volatiles, significant quantities of these species exist within the Moon.

Moderately volatile elements (MVEs) are defined as elements with a condensation temperature between 650-1350K [7]. Their higher condensation temperatures potentially allows them to record condensation/degassing events [7]. The concentration of MVEs in lunar basalts provide insight not only into the overall level of degassing the Moon experienced, but also to the MVE inventory in the lunar mantle.

Mare basalts are the result of partial melting of compositionally distinct regions of the cumulate lunar mantle [8,9]. The material that coalesced after the Giant Impact would have been mostly, if not entirely, molten. The subsequent crystallization can be thermodynamically modeled and would have mineralogically followed Bowen’s reaction series [9]. Forsteritic Olivine (OL) would have formed first, and being denser than the surrounding melt, would have settled to the base of the lunar magma ocean (LMO) [9]. As crystallization progressed from OL to orthopyroxene (OPX) to clinopyroxene (CPX), each phase would be progressively more Fe-rich until ilmenite (ILM) came on the liquidus [9]. The last dregs of the LMO would be enriched in incompatible elements and has been termed urKREEP [10]. As the mantle cumulate formed, the Fe-rich phases would be stratigraphically above the Mg-rich phases. At some point after the crystallization of dense ILM, the mantle cumulate became gravitationally unstable and went through an overturn process that mixed to varying degrees the early and late cumulates [9,11]. The efficiency of the turnover and the resulting heterogeneity of the lunar mantle is still debated. Mare basalts are

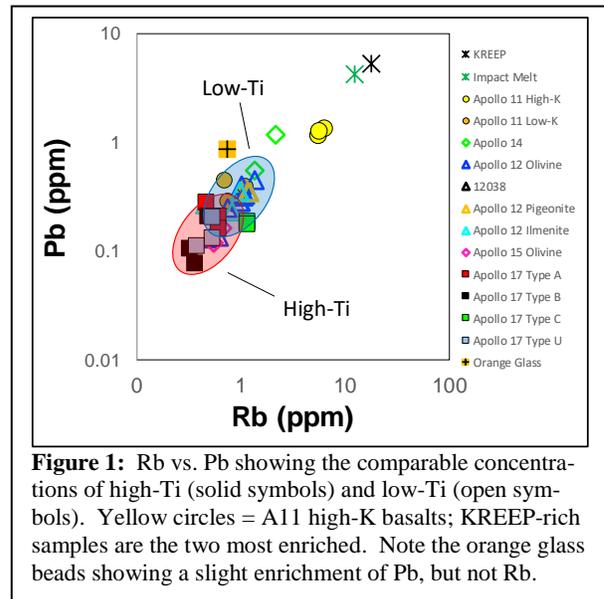


Figure 1: Rb vs. Pb showing the comparable concentrations of high-Ti (solid symbols) and low-Ti (open symbols). Yellow circles = A11 high-K basalts; KREEP-rich samples are the two most enriched. Note the orange glass beads showing a slight enrichment of Pb, but not Rb.

categorized as very-low-Ti (VLT), low-Ti, and high-Ti (e.g., [12]). VLT and low-Ti represent partial melts of a source region dominated by early LMO cumulates, whereas high-Ti basalts were derived from a source dominated by late-stage LMO cumulates. KREEP basalts contain a significant urKREEP component and are derived from source regions with a lower Ca/Al ratio than mare basalts [8].

Methods: The MVEs chosen for this study are Zn, Se, Rb, Ag, Cd, In, Sb, Tl, Pb, & Bi. They have been measured in 54 whole rock samples using solution mode inductively coupled plasma mass spectrometry (SM-ICP-MS). The trace nature of these MVEs makes them especially susceptible to interferences that occur when another element in the sample forms a polyatomic ion that equals the element of interest mass and elevates the signal. To mitigate this, a series of interference solutions were designed to quantify and correct for the data [13]. Standard reference materials (SRMs) BIR and BHVO were run in conjunction with the lunar samples to ensure analytical accuracy.

Results & Discussion: The generally incompatible nature of these MVEs with the crystallizing phases in the LMO means that as crystallization progressed, they should be progressively more enriched in the remaining melt. The only exception to this is Zn which has shown to be compatible in both OPX and ILM [14]. The resulting concentrations should show a positive trend between

the low-Ti, high-Ti, and KREEP basalts. The Apollo 11 high-Ti, high-K samples are more enriched due to the incorporation of a KREEP-rich component [15] (Fig 1). An exception to this is Zn where these high-K samples have comparably high Ag to high-Ti basalt concentrations, but plot below the low-Ti basalts with respect to Zn (Fig 2). The compatible nature of Zn in ILM and OPX may explain the slight overlap in concentrations seen in the low and high-Ti samples (Fig 2). The Apollo 12 ilmenite basalts, while traditionally categorized as low-Ti basalts, have been shown to have an intermediate composition and span the concentrations of both the low and high-Ti basalts (e.g., [16]). A KREEP-rich impact melt (14310) was analyzed and although it was not produced by mantle melting processes, shows similar enrichments as pristine KREEP basalts. The data show that the expected enrichment between the low- and high-Ti basalts is not present with the exception of Ag (Fig 2) and Se, which may be due to their more chalcophile behavior. If the LMO underwent sulfide immiscibility, this dense S-rich phase may have sunk, percolating through and metasomatizing the earlier cumulates or unmelted region of the Moon [17]. An event such as this would have affected highly chalcophile Ag and Se relative to the other elements in the MVE suite.

Orange volcanic glass beads, shown to be some of the most volatile rich material collected on the Moon [6], were also hand-picked and analyzed. These data show the orange glass is enriched in the MVEs relative to mare basalts, but not uniformly across all analyzed elements. The orange glass is KREEP-like with respect to Zn & Ag (Fig 2), but similar to the low-basalts with respect to Rb & only slightly enriched in Pb (Fig 1).

Like the MVEs, Ti is incompatible in the major LMO phases, except ILM, but it is more refractory. Normalizing these data to Ti should allow direct

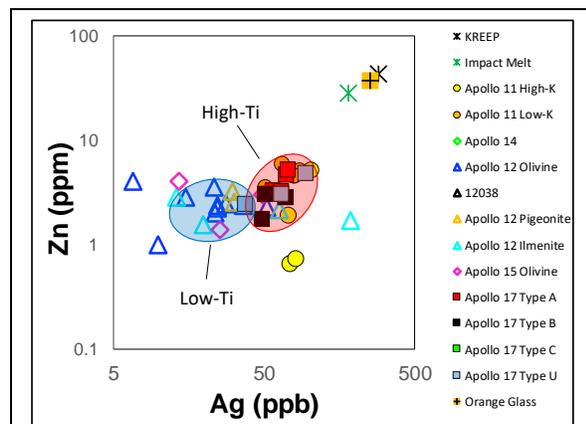


Figure 2: Ag vs. Zn showing the variation in Zn concentrations in A11 high-Ti samples. This could be due to ilmenite fractionation, or degassing during eruption.

comparisons between the low- and high-Ti samples. The high-Ti samples are depleted in MVEs relative to the low-Ti basalts (Fig 3) which is interpreted to represent the degassing of MVEs from the LMO after the formation of the low-Ti basalt source. This could be the result or combination of several processes; steady LMO convection restricting the formation of a volatile-sealing quenched lid [18]; an impact large enough to breach the crust and degass the late stage cumulate volatile reservoir [19] followed by resealing. We prefer the latter option on the basis of our data.

Conclusion: The MVE content of a suite of mare and KREEP basalts have been quantified. The depletion in high-Ti basalts relative to low-Ti basalts is interpreted as an overall loss in volatile species in the late-stage LMO cumulates, contrary to expectations. This has been interpreted as a LMO degassing event that affected the later stage, high-Ti cumulates, probably through a crust-penetrating impact.

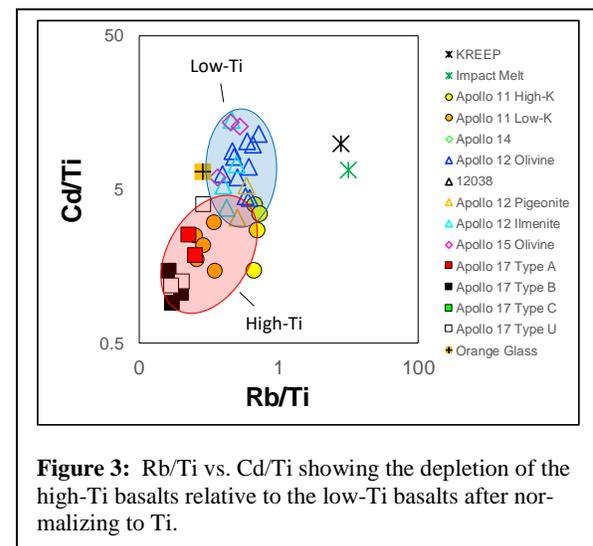


Figure 3: Rb/Ti vs. Cd/Ti showing the depletion of the high-Ti basalts relative to the low-Ti basalts after normalizing to Ti.

References: [1] Hartmann W.K. & Davis D.R. (1975) *Icarus* 24, 504-515. [2] Canup, R.M. & Asphaug, E. (2001) *Nature* 412, 708-712. [3] Anders, E. (1970) *Science* 169, 1309-1310. [4] Hauri E. et al. (2011) *Science* 333, 213-215. [5] McCubbin F. et al. (2015) *Amer. Mineral.* 100, 1668-1707. [6] Saal A. et al. (2008) *Nature* 454, 192-195. [7] Palme, H. et al. (1988) *Meteorites and the Early Solar System I*, 436-461. [8] Hallis, L.J. et al. (2014) *GCA* 134, 289-316. [9] Elkins-Tanton, L.T. et al. (2011) *EPSL* 304, 326-336. [10] Warren P.H. (1985) *AREPS* 13, 201-240. [11] Spera F. (1992) *GCA* 56, 2253-2265. [12] Neal C.R. & Taylor L.A. (1992) *GCA* 56, 2177-2211. [13] Burney, D.C. et al. (2017) *LPSC* 48, #1068. [14] Klemme, S. et al. (2006) *Chem. Geo.* 234, 251-263. [15] Jerde E.A. et al. (1994) *GCA* 58, 515-527. [16] Snyder G.A. et al. (1997) *GCA* 61, 2731-2747. [17] Neal, C.R. (2001) *JGR* 106(E11), 27865-27885. [18] Dhaliwal, J.K. et al. (2018) *Icarus* 300. [19] Barnes et al. (2016) *EPSL* 447, 84-94.