

RE-EXAMINING THE PETROGENESIS OF LUNAR TROCTOLITES. T.C. Prissel^{1,2}, J. Gross^{1,2}. ¹Department of Earth & Planetary Sciences, Rutgers University, Piscataway, NJ 08854. ²American Museum of Natural History, New York, NY 10024. (contact: tcprissel@gmail.com).

Background: Lunar highlands magnesian-suite samples (Mg-suite) are comprised of pristine, plutonic igneous rock fragments and clasts that represent the earliest stages of lunar magmatism (~ 4.3 Ga) [e.g., 1-5]. In particular, primitive end-member Mg-suite troctolites and volumetrically minor pink spinel troctolites (PST) contain the most forsteritic olivine compositions ($Fo\# = Mg\# = Mg/(Mg + Fe) \times 100 =$ up to Fo96) among all lunar samples (Fig. 1), suggesting the Mg-suite parent magma is derived, at least in part, from ultramafic cumulates produced during primary differentiation of a global lunar magma ocean (LMO; e.g., 6-10).

Premise: Apollo Mg-suite samples are interpreted to be co-magmatic: a common parental magma first produced primitive (higher Mg/Fe) olivine-rich lithologies including dunites, troctolites, and PST, with continued fractionation of the parent magma resulting in more evolved (lower Mg/Fe) pyroxene-rich daughter lithologies such as norites and gabbro-norites [e.g., 1,4,11]. Regarding primitive lunar troctolites and rarer PST, the unique pairing of forsteritic olivine with anorthitic-plagioclase ($An\# = Ca/(Ca + Na + K) \times 100 < 97$) has been used to argue against simple models of equilibrium crystallization from typical mantle-derived magmas. The reason being typical mantle-derived magmas are plagioclase-undersaturated and require a substantial amount of olivine fractionation (> 45%) before crystallizing plagioclase; a process that will lower the Mg/Fe of the magmatic system, co-crystallizing relatively FeO-rich olivine with plagioclase. For instance, [12] demonstrates that a komatiitic basalt would be co-saturated with Fo86 olivine and An97 plagioclase after ~45% olivine crystallization. The Fo + An compositions produced in the example of [12] are within the compositional range observed among lunar troctolites (Fig. 1), but contain lower Fo# than a majority of lunar troctolites, and would be ~10 Mg# units below the most forsteritic olivine in PST (Fo96, see 13). These results have been used to define the so-called “Mg# problem” associated with the formation of lunar troctolites by equilibrium crystallization from plagioclase-undersaturated magmas [12]. Alternative endogenic models have explored the possible presence of a plagioclase-bearing hybridized source region (capable of producing plagioclase-saturated parental melts) [e.g., 7,14], and the production of plagioclase-saturated melts via magma-wallrock interactions within the anorthositic crust [e.g., 12,15].

Recently, the experimental study of [9] demonstrated that Mg-suite parental melts near plagioclase-saturation

prior to crystallization exclusively yield pink spinel – indicating only the rarer PST can be the result of magma-wallrock interaction or a plagioclase-bearing hybridized source region. The more common troctolites (+/- chromian spinel) do not contain pink spinel, and therefore may have formed via equilibrium crystallization from a plagioclase-undersaturated, MgO-rich mantle-derived magma [9]. Note that only the PST contain $Fo\# > 90$, and common lunar troctolites (+/- chromian spinel) contain $Fo\# \leq 89$ with a majority at $Fo\# \sim 87-88$ (Fig. 1). When distinguishing common troctolites from PST, the lower overall Fo# associated with common troctolites appears consistent with the equilibrium crystallization model of [12] (Fig. 1). Specifically, the troctolite assemblage of [12] (Fo86+An97) is only 3 Mg# units below the most forsteritic olivine among common troctolites (Fig. 1), suggesting the Mg-suite parent need only be slightly more MgO-rich than the specific komatiitic basalt modeled in [12].

Equilibrium Crystallization of Plagioclase-Undersaturated Magmas: We have modeled the crystallization sequence and resulting troctolite mineralogy of estimated Mg-suite parent magmas using the SPICEs (Simulating Planetary Igneous Crystallization Environments) matlab equilibrium crystallization code [17]. Below, we test the komatiitic basalt composition of [12] to ensure consistency with previous results, and compare with estimates from [15]. These results are then used to inform new estimates of Mg-suite parental magmas [16].

The komatiitic basalt from [12] produces a troctolite assemblage of Fo85.8 olivine, En84.7 orthopyroxene, and An97.9 plagioclase after ~45% olivine crystallization (compared to Fo86.2 + An97 reported by 12). The resulting troctolite assemblage modeled here is consistent with [12], noting the addition of co-precipitating orthopyroxene ~57°C prior to plagioclase saturation. A small modal abundance of orthopyroxene (typically < 10%) exists in many lunar troctolites with a limited compositional range from En85-91 and En70-91 in the common troctolites and PST, respectively [4].

The “standard initial” (SI) composition from [15] yields Fo83.7, En83.1, and An98.3. The resulting troctolite assemblage is consistent with the lowest Mg# troctolites in the current sample database, and ~2 Mg# units less than the troctolite assemblage from [12].

Model results indicate that only a slightly more MgO-rich komatiitic basalt is needed to explain the formation of the most forsteritic common troctolites ($Fo < 89$) via equilibrium crystallization. Expectedly similar results are

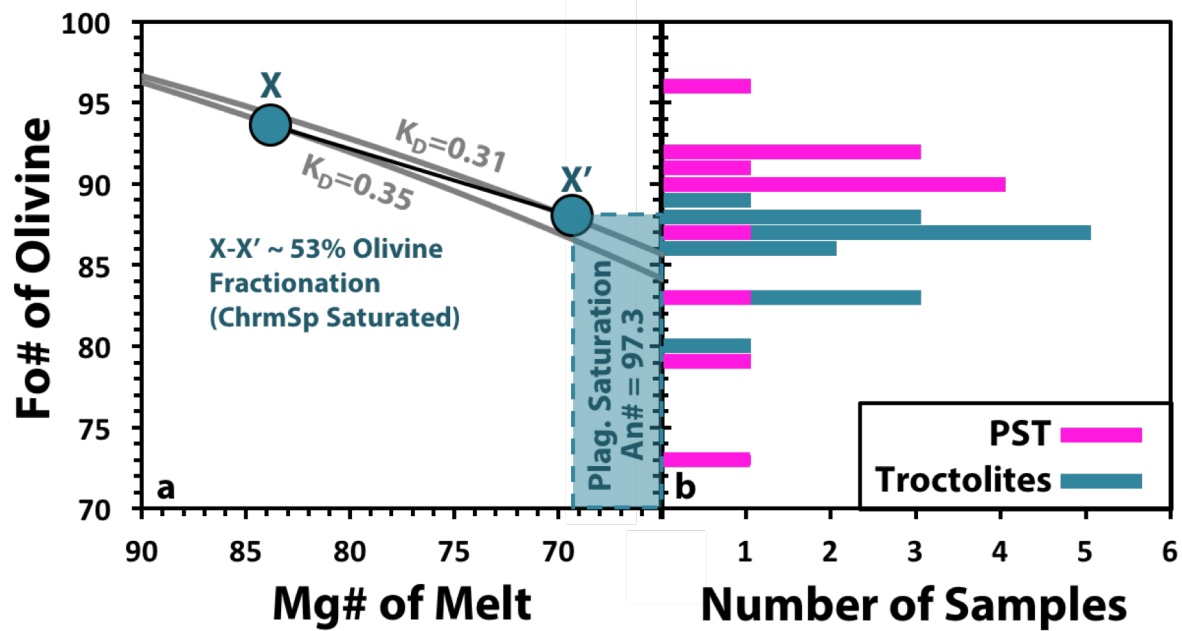


Fig. 1. From [16]. The forsterite content (Fo# = Mg#) of olivine as a function of (a) the Mg# of a crystallizing melt, and relative to (b) the total number of Mg-suite samples. Magmas in equilibrium with Fo# 94 liquidus olivine (X) become multiply-saturated after ~53% olivine (+ trace chromian spinel) crystallization (X') - yielding Fo89 olivine, and An97 plagioclase. Note that only pink spinel troctolites (PST) contain olivine with Fo# > 89. Thus, the mineralogy produced after 50% olivine fractionation is most consistent with common lunar troctolites (+/- chromian spinel), suggesting formation by equilibrium crystallization of a plagioclase-undersaturated magma [9].

produced when modeling the crystallization sequence of a komatiitic basalt in equilibrium with liquidus Fo94 (Fig. 1). The more MgO-rich komatiitic basalt modeled here yields a troctolite assemblage of Fo89.1 and An97.5 (+ chromian spinel) after ~53% olivine fractionation. The resulting troctolite assemblage is consistent with the most forsteritic common troctolites.

Summary & Future Work: In light of experimental constraints from [9], we have demonstrated that equilibrium crystallization of a MgO-rich komatiitic basalt yields mineralogy consistent with the most forsteritic common troctolite samples. Such parental magmas are expected from melting of primordial ultramafic cumulates (possibly pressure-release melting during lunar mantle overturn) [4,9,12,15,18,19]. Results are consistent with earlier models of equilibrium crystallization [e.g., 20], albeit with a more MgO-rich parent.

Several petrogenetic consequences follow if common troctolites formed during equilibrium crystallization of a plagioclase-undersaturated magma: (I) results preclude the geophysical processes required to produce a hypothesized plagioclase-rich Mg-suite hybridized mantle reservoir. We conclude the formation of common troctolites can be attributed to equilibrium crystallization of MgO-rich komatiite-like mantle melts [12,19], whereas PST mineralogy can be explained as products of the same magma interacting with anorthositic crust [9,16]. Further, (II) a substantial amount of latent heat would be released

during > 45% forsteritic olivine crystallization. Previously, the latent heat of crystallization was considered to be a minor-to-non-factor in crustal assimilation models on the basis of the “Mg# problem.” This has prompted our thermodynamic reevaluation of crustal assimilation [16], which will be presented in conjunction with this work. On a larger-scale, (III) further constraining the bulk composition of the Mg-suite parent will directly inform our chemical understanding of the deep lunar interior as well as geophysical models of LMO differentiation.

References: [1] James O.B., (1980) *11th Proc. LPSC* 365-393 [2] Nyquist L.E. & Shih C.Y. (1992) *GCA* 56, 2213-2234 [3] Papike J.J. et al. (1998) *RMG* 36 [4] Shearer C.K. et al. (2015) *Am. Min.* 100, 294-325 [5] Borg L.E. et al. (2016) *GCA* 201, 377-391 [6] Snyder G.A. et al. (1992) *GCA* 56, 3809-3823 [7] Elardo S.M. et al. (2011) *GCA* 75, 3024-3045 [8] Elkins-Tanton L.T. et al. (2011) *EPSL* 134, 501-514 [9] Prissel T.C. et al. (2016) *Am. Min.* 101, 1624-1635 [10] Lin Y. et al. (2017) *EPSL* 471, 104-116 [11] Norman M. & Ryder G. (1979) *Proc. 10th LPSC*, 531-559 [12] Hess P.C. (1994) *JGR* 99, 19083-19093 [13] Snyder G.A. et al. (1999) *30th LPSC* #1491 [14] Longhi, J. et al. (2010) *GCA* 74 784-798 [15] Warren P.H. (1986) *JGR* 91, D331-D343 [16] Prissel T.C. & Gross J. (in review) [17] Davenport J.D. et al. (2014) *45th LPSC* #1111 [18] Hess P.C. & Parmentier E.M. (1995) *EPSL* 134, 501-514 [19] Shervais J.W. & McGee J.J. (1998) *GCA* 62, 3009-3023 [20] Snyder G.A. et al. (1995) *JGR* 100, 9365-9388