

## Fe<sup>2+</sup>-Mg PARTITIONING BETWEEN OLIVINE AND MARTIAN BASALTIC MELTS.

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**Introduction:** The partitioning of elements between olivine and liquid is key to understanding igneous processes in mafic and ultramafic systems [1]. The landmark study of Roeder and Emslie [2] concluded that the olivine (ol)-liquid (liq) exchange coefficient,  $K_{D,Fe^{2+}-Mg} = (FeO/MgO)^{ol} / (FeO/MgO)^{liq}$  (by weight), is  $0.30 \pm 0.03$ , independent of temperature ( $T$ ) and liquid composition. Subsequent work [3] has shown a dependence on liquid composition, but the canonical value of 0.30 is still widely used in both terrestrial and Martian applications [e.g., 4, 5]. Here, we use previously published experiments on bulk compositions relevant to Martian magmas to determine the appropriate  $K_{D,Fe^{2+}-Mg}(S)$  for Martian systems; we then use this value to test whether any of the ol-phyric shergottites represent liquids.

**Experiments on Martian Compositions:** One-atm experiments on model Martian compositions provide the tightest constraint on  $K_{D,Fe^{2+}-Mg}$  because  $T$  and oxygen fugacity ( $fO_2$ ) are well known, allowing us to use [6] to predict the amount of ferrous iron present in each liquid. The median of 17 published 1-atm experiments yields  $K_{D,Fe^{2+}-Mg} = 0.354 \pm 0.008$  (error is one mean absolute deviation, MAD). There are higher pressure ( $P$ ) experiments on model Martian compositions, but their  $fO_2$ s are less well constrained. High- $P$  experiments in graphite capsules are reducing [7], but  $Fe^{3+}$  still materially affects estimates of  $K_{D,Fe^{2+}-Mg}$ ; correcting for the  $Fe^{3+}$  using [6,7] increases  $K_{D,Fe^{2+}-Mg}$ , on average, by 0.016. We see no strong correlations of  $K_{D,Fe^{2+}-Mg}$  with liquid or ol composition for the high or low- $P$  experiments and a weak increase in  $K_{D,Fe^{2+}-Mg}$  with increasing  $P$ . A marked increase in  $K_{D,Fe^{2+}-Mg}$  for high- $P$  experiments at  $T < \sim 1150^\circ C$  is likely due to lack of equilibrium. Accepting high- $P$  experiments with  $T > 1150^\circ C$  leads to a median  $K_{D,Fe^{2+}-Mg}$  of  $0.369 \pm 0.024$  (MAD), higher than the 1-atm experiments but notably higher than the canonical value of 0.30 [2]. Combining 1-atm. and high- $P$ , high- $T$  experiments yields a median  $K_{D,Fe^{2+}-Mg}$  of  $0.360 \pm 0.013$  (MAD), close to the recently reported value of  $0.35 \pm 0.01$  [8], which was obtained without correcting for  $Fe^{3+}$  in the liquid.

**Olivine-phyric Shergottites:** Using  $K_{D,Fe^{2+}-Mg} = 0.30$  leads to the result that none of the ol cores in ol-phyric shergottites are in equilibrium with liquids equivalent to their bulk. Applying our  $K_{D,Fe^{2+}-Mg}$  to these shergottites, leads to the possibility that Y980459, NWA 5789 and 2990 are liquid compositions (others are not), identical to the results of [8]. Accounting for  $Fe^{3+}$  in the liquid also allows us to constrain plausible  $fO_2$ s during cooling when independent measures are unavailable or not yet determined; e.g., if the olivines in NWA 2990 are in equilibrium with a liquid whose composition is that of the bulk meteorite, crystallization must have occurred under reducing conditions where the  $Fe^{3+}/\Sigma Fe$  is small,  $\sim IW+0.5$  at 1 atm or  $IW+1$  at 1 GPa.

**References:** [1] Basaltic Volcanism Study Project. 1981. [2] Roeder P. L. and Emslie R. F. 1970. *Contrib. Mineral. Petrol.* 29:275–289. [3] Sack R. O. et al. 1987. *Contrib. Mineral. Petrol.* 96:1–23. [4] Falloon T. J. et al. 2007. *Chem. Geol.* 241:207–233. [5] Peslier A. H. et al. 2010. *GCA* 74:4543–4576. [6] O'Neill et al. 2006. *Am. Mineral.* 91:404–412. [7] Médard E. et al. 2008. *Am. Min.* 93:1838–1844. [8] Filiberto J. and Dasgupta R. 2011. *EPSL* 304:527–537.