

Fe²⁺-Mg PARTITIONING BETWEEN OLIVINE AND MARTIAN BASALTIC MELTS.

A. K. Matzen, J. R. Beckett, M. B. Baker, and E. M. Stolper.
California Institute of Technology, MC 170-25, Pasadena, CA
91125. amatzen@caltech.edu.

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 ± 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 ± 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 ± 0.013 (MAD), close to the recently reported value of 0.35 ± 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.