Qiong-Yao Zhan, Di-Cheng Zhu, Roberto F. Weinberg, Qing Wang, Jin-Cheng Xie, Liang-Liang Zhang, Zhi-Dan Zhao. Cumulate granites: A perspective from new apatite MgO partition coefficients: *Geology*, <https://doi.org/10.1130/G49781.1>

Supplemental Material

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1. Data processing methods

1.1. Experimental apatites and melts

Compositions of apatite-melt pairs are compiled from 184 experimental runs conducted in 14 phase equilibrium experiments (Table S1). Individual experimental runs were conducted at different conditions (e.g., composition, temperature, and pressure). Each run contains several analyses of melts and apatite grains and the average compositions of melts and apatites with corresponding standard deviations (SD) are available. Apatite and melt MgO and other common oxides (e.g., SiO₂, FeO, and CaO) are all analyzed by EPMA. Before using these experimental data to explore the relationship between apatite MgO and coexisting melt MgO, it is necessary to determine whether the apatite and melt are in equilibrium for the composition of MgO and whether the analysis are of high quality. For an equilibrium experimental run, both melts and different apatite grains are expected to be homogenous in composition and the SD of MgO are only related to analytical methods. For an experimental run that is not in equilibrium, either melts or different apatite grains are expected to be heterogeneous in composition and the SD of MgO are related to not only analytical methods but also the inhomogeneity. Therefore, for experimental runs that are not in equilibrium or associated with poor analysis, the melts or apatites or both are expected to have large relative standard deviations (RSD) of MgO.

A plot of RSD of apatite and melt MgO versus their MgO contents (Fig. S1A) for the 184 experimental runs shows that the RSD values are generally <20% for MgO >~0.6 wt% but shows large variations when MgO decreases below ~0.6 wt%. Therefore, experimental runs with RSD of MgO >20% are excluded for further consideration. The detection limit of MgO by EPMA of apatites are not available for these studies except for one experimental study that reported the detection as 0.03 wt% (Konecke et al., 2017). Therefore, data with apatite MgO <0.03 wt% are excluded. The experimental runs fall into three trends by plotting the apatite MgO against equilibrium melt MgO (Fig. S1B). Trend 3 is somewhat special because these experimental runs were conducted at low oxygen fugacity at ~IW buffer (Fig. S1C) and comprise starting melt composition with very high FeO content (>12 wt%, Fig. S1D), mimicking Martian meteorites, and low CaO content, resulting in strongly peraluminous basaltic melt (Fig. S1E, McCubbin et al., 2015, 2018). These experimental conditions result in apatites with high FeO contents (1.91–5.11 wt%) as well as high MgO contents (>0.7 wt%), which is unusual compared with typical

apatites (Fig. S1F). Therefore, trend 3 describes poorly the apatite-melt equilibrium relationship in natural systems and is not shown in Fig. 1A in the main text to avoid confusing.

1.2. Volcanic and plutonic apatites, glass, and whole rocks

Compositions of volcanic and plutonic apatites, glass, and whole rocks (Table S2–S4) are synthesized based on the precompiled dataset of apatites in the GEOROC database (by 10/13/2021). Apatite and glass MgO and other common oxides (e.g., SiO₂, FeO, and CaO) are all analyzed by EPMA. Whole-rock major elements of volcanic and plutonic rocks are analyzed by either XRF or wet chemistry.

For the data set of volcanic apatites and glass, only apatites clearly stated as occurring in glass or in matrix are included. Apatites as inclusions in other minerals or with unclear occurrence are not included. Compositions of apatites and coexisting glass are available for 39 volcanic rocks (Table S2). Although it is difficult to prove the equilibrium between volcanic apatite and coexisting glass, some disequilibrium situations can be determined based on the large RSD of either melt MgO or apatite MgO. Following the criteria used for experimental data, volcanic apatite-melt pairs with RSD of MgO >20% or MgO < 0.03wt% are excluded. In addition, among these volcanic rocks, three samples have glass showing large compositional variations as stated in the literature (Peretyazhko et al, 2018), but only average compositions of apatites and glass are reported and SD are not available. Therefore, these three samples are also excluded. Two samples of Timina et al (2006) yield low total oxides of glass compositions (total < 93wt%) suggesting poor analysis or alteration. Therefore, these two samples are also excluded. After this filtering process, 7 samples are kept and are shown in Fig. 1 in the main text.

For the data sets of volcanic and plutonic apatites and whole rocks, some rock types including carbonatites, kimberlites and lamprophyres are not incorporated in our compilation. This is because one of the primary goal of this study is using apatite to explore the relationships between volcanic rocks and plutonic rocks, in particular whether granitic rocks are cumulates. No filtering processes are applied to volcanic and plutonic data although some apatites in plutonic rocks have MgO below the detection limit. This is because the exact value is not so important here and what important is that apatites in some plutonic rocks have very low MgO (Fig. S2), suggesting that the apatites crystallized from evolved melt composition rather than the host bulk rock composition.

2. Controls on Mg partitioning between apatite and melt

Several factors may control element partition between apatite and melt include pressure (P), temperature (T), oxygen fugacity (f_{O_2}), crystal composition, and melt composition or polymerization (e.g., [Prowatke and Klemme, 2006](#); [Stokes et al., 2019](#)). Partition coefficient of Mg between apatite and melt ($D_{Mg}^{Ap/Melt}$, mathematically equals $D_{MgO}^{Ap/Melt} = MgO_{Ap}/MgO_{Melt}$) can be calculated from each experimental run ([Table S1](#); single point in [Fig. 1A](#) in the main text). Exploring whether $D_{MgO}^{Ap/Melt}$ correlates with potential controlling factors can provide valuable insights into the controls on $D_{Mg}^{Ap/Melt}$.

It turns out that $D_{MgO}^{Ap/Melt}$ does not show correlation with P , T , and f_{O_2} ([Fig. S3](#), see [Table S1](#) for these parameters), suggesting that these factors have no or limited controls on $D_{Mg}^{Ap/Melt}$. Possible effect of crystal compositions on $D_{Mg}^{Ap/Melt}$ is evaluated in terms of apatite halogens that are F and Cl. There are no correlations between $D_{MgO}^{Ap/Melt}$ and halogens ([Fig. S3](#)), indicating that halogens have no controls on $D_{Mg}^{Ap/Melt}$. The effects of melt composition on $D_{Mg}^{Ap/Melt}$ can be evaluated in terms of alumina saturation indices ($A/CNK = \text{molar } Al_2O_3/(CaO+Na_2O+K_2O)$) or polymerization characterized by the ratio of non-bridging oxygens (NBO) to tetrahedrally coordinated cations (T). There is no correlation between $D_{MgO}^{Ap/Melt}$ and A/CNK ([Fig. S3](#)). In contrast, $D_{MgO}^{Ap/Melt}$ correlates with melt NBO/T in an exponential function, suggesting that melt polymerization exerts a strong control on the $D_{Mg}^{Ap/Melt}$ (see discussion in the main text). NBO/T values are calculated as follows:

- (1) Convert the chemical analysis of the melt, glass, or rock to atomic proportions.
- (2) Assign Si, Ti, and P to T.
- (3) Assign Al to T if $Al < Na + K + 2Ca + 2Mg$. If $Al > Na + K + 2Ca + 2Mg$, Al in T = $Na + K + 2Ca + 2Mg$.
- (4) If $(Na + K) - Al < 0$, Fe in T = 0. If $(Na + K) - Al > 0$ but $< Fe^{3+}$, Fe in T = $(Na + K) - Al$. If $(Na + K) - Al > 0$ and $> Fe^{3+}$, Fe in T = Fe^{3+} .
- (5) Sum all T cations = T.
- (6) $NBO = 2O - 4T$.
- (7) Then calculate NBO/T.

In calculation, ferric/ferrous iron ratios of experimental melts are estimated using the experimental calibration of [Borisov et al \(2018\)](#). Proportions of Fe_2O_3 and FeO of volcanic glass were estimated from total iron using ferric/ferrous iron ratios suggested by [Middlemost \(1989\)](#)

for each volcanic rock type as classified on the total alkalis versus silica diagram. Calculation processes are given in [Table S5](#). See also the spreadsheet of [Prata et al \(2019\)](#) for an example of calculation.

3. On the diffusion of Mg in apatite

Mg diffusion behavior in apatite is currently unconstrained. There are studies that explore the diffusion behavior of other divalent cations in apatite, such as Mn, Sr, Pb, and Ca (see [Cherniak, 2010](#) for a review). The diffusion rates of Mn, Sr, Pb, Ca is in the order of $Mn < Sr < Pb < Ca$. The diffusion rates decrease with decreasing ionic radius except for Ca. Mg^{2+} has an ionic radius smaller than Mn^{2+} . It may deduce that the diffusion rate of Mg is similar to or lower than Mn. The diffusion rate of Mn in apatite is low ([Fig. S4A](#)). This is supported by our compiled plutonic dataset which shows that plutonic apatite MnO shows large variation ([Fig. S4B](#)). If Mn in apatite diffuses fast, MnO in apatite is expected to re-equilibrated easily and low MnO rocks will have low MnO apatites, which is contrast to the plutonic apatites. Therefore, the diffusion rate of Mg could be also slow. Further experimental studies are needed to constrain the diffusion behavior of Mg in apatite.

4. Datasets and figures

Tables S1–S5 are listed in Excel files.

Table S1. Compositions of apatites and coexisting melts from phase equilibrium experiments

Table S2. Compositions of apatites and coexisting glass in volcanic rocks

Table S3. Compositions of volcanic apatites and host rocks

Table S4. Compositions of plutonic apatites and host rocks

Table S5. Calculations of experimental melt and volcanic glass NBO/T

Figure S1. (A) Relative standard deviations (RSD) of MgO plotted as a function of MgO contents for experimental melts and apatites. (B) A plot similar to Fig. 1A in the main text but with filtered data also shown. (C–D) Temperature, oxygen fugacity, and melt FeO (total Fe) of experimental runs in the three trends. Some experimental runs have no oxygen fugacity available or their FeO contents are below detection limit. Therefore, these experimental runs are not shown. (E) Experimental melt A/CNK (molar ratio of $\text{Al}_2\text{O}_3/(\text{CaO}+\text{Na}_2\text{O}+\text{K}_2\text{O})$) vs melt SiO_2 . (F) FeO vs MgO for experimental and natural apatites. Experimental data sources are WG81 (Watson and Green, 1981), R91 (Ryabchikov et al., 1991), MW05 (Mathez and Webster, 2005), PK06 (Prowatke and Klemme, 2006), K17 (Konecke et al., 2017), R18 (Riker et al., 2018), and S19 (Stokes et al., 2019) for trend 1, and W09 (Webster et al., 2009), D14 (Doherty et al., 2014), LH15 (Li and Hermann, 2015), LH17 (Li and Hermann, 2017), W17 (Webster et al., 2017), and S19 (Stokes et al., 2019) for trend 2, and M15 (McCubbin et al., 2015) and MU18 (McCubbin and Ustunisik, 2018) for trend 3.

Figure S2. (A–B) A same plot as [Fig. 2](#) in the main text but with x- and y-axis at log scale. Granitic rocks with high apatite saturation temperatures ($AST > 900^{\circ}C$) and low $AST (< 800^{\circ}C)$ are highlighted. AST is calculated following the method of [Harrison and Watson \(1984\)](#) for metaluminous rocks and [Bea et al \(1992\)](#) for peraluminous rocks. (C–D) A same plot as (A–B) but the volcanic and plutonic rocks are both divided into two groups based on the tectonic setting including convergent margin and intraplate setting.

Figure S3. $D_{\text{Ap/Melt}}^{\text{MgO}}$ plotted against experimental pressure (P), temperature (T), oxygen fugacity (f_{O_2}), apatite halogens, and melt alumina saturation indices (A/CNK = molar $\text{Al}_2\text{O}_3/(\text{CaO}+\text{Na}_2\text{O}+\text{K}_2\text{O})$). Log oxygen fugacity is calculated from experimental redox buffers at the experimental temperature using the relevant redox buffer equations compiled by Frost (2018).

Figure S4. (A) Plot of diffusion distance of Mn (dMn) per million year ($\mu m/Ma$) vs temperature showing that Mn diffusion in apatite is not fast. dMn is calculated from the volume diffusion rate of Mn (DMn , m^2/s) which can be calculated using equation reported by [Cherniak \(2010\)](#) and $dMn = (DMn)^{1/2}$ ([Zhang et al., 2010](#)). Three dMn values are marked beside data points which have $T < 800$ $^{\circ}C$. (B) Plutonic apatite MnO vs host rock MnO further showing that Mn diffusion in apatite is slow. If Mn diffuses fast, MnO in apatite is expected to re-equilibrate easily and therefore, plutonic rocks with low MnO will have low MnO apatites. (C) REE patterns and (D) ternary plot of F, Cl, and OH atomic proportions for apatites in our compiled granitic rocks which have high apatite saturation temperatures ($AST > 900$ $^{\circ}C$). F and Cl are determined by EPMA and OH was estimated following the method of [Ketcham \(2015\)](#).

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