From melt- to crystal-rich magmatic systems during rift localization:
Insights from mineral chemistry in Central Afar (Ethiopia)

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Supplementary Text

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Olivine mineral chemistry results

The US olivine have a wider compositional range (range $Mg\#_{29-85}$) with respect to the CAG (range $Mg\#_{55-82}$) and a bimodal distribution of core compositions ($Mg\#_{74}$ and $Mg\#_{83}$) while the CAG have a unimodal distribution ($Mg\#_{77}$) (Fig. S6A, B, C, D). Although the US core bimodality is not systematically related to their provenance from low-MgO or high-MgO rock groups, some variations between them have been observed (Fig. S6E, F, G, H). Iddingsite is present at the rim and along the fractures of some crystals, and were avoided as far as possible during the analysis.

Regarding the US, the high-MgO samples have olivine with normal zoning or unzoned with cores in the range $Mg\#_{69-84}$ and modest intrasample variation (Fig. S6E, the bimodal distribution is related to distinct samples). The US low-MgO samples have few and normally zoned olivine with slightly lower Mg# with respect to the high MgO samples (Fig. S6E, G). Regarding the CAG the high-MgO samples have olivine with normal zoning or unzoned with core composition having the higher Mg# values recorded among the CAG (range $Mg\#_{76-82}$; Fig. S6F). The low-MgO samples have olivine with normal zoning, with core composition corresponding to the lower Mg# values of the CAG (range $Mg\#_{74-80}$; Fig. S6H).

Fe-Ti oxides mineral chemistry results

Fe-Ti oxides have been observed as accessory minerals in all samples, sometimes enclosed at the rim or mantle of phenocrysts (Fig. S1). Both the US high and low-MgO samples have ilmenite and titanomagnetite with ulvöspinel component up to 78% (Fig. S7). No clear coexisting pairs have been found. Both CAG high and low-MgO samples have ilmenite (except for one low-MgO sample) and titanomagnetite (Fig. S7). Furthermore, chromite has been observed in both the US and CAG high-MgO samples, always enclosed in olivine (Tab. S1).

Mineral-melt equilibrium and geothermobarometry results

Geo-thermobarometric estimations have been obtained using the mineral compositions of plagioclases and clinopyroxenes in both the two Series. We used the plagioclase geothermobarometer of Putirka (2008) to calculate the pressure (eq. 25a, standard errors of estimate (SEE) = ±3.8 kbar) and temperature (eq. 24a, SEE = ±36 °C) of crystallization. The clinopyroxene pressure and temperature of crystallization has been calculated using the methods of

Neave and Putirka (2017; SEE = ±1.4 kbar) and Putirka (2008; eq. 33, SEE = ±45 °C) respectively (Tab. S1). We used the whole rock analysis as the melt composition, according to Putirka (2008). We used a water concentration input of 0.5 wt% according to evidence from both literature of low water content of Afar mafic magmas (Field et al. 2012, 2013), and MELTS best fit model of Tortelli et al. (2022). The equilibrium was checked based on the distribution coefficient K_D(An-Ab)pl−liq = 0.27 ± 0.11 at T ≥ 1050°C for the high- and low-MgO samples plagioclase, K_D(An-Ab)pl−liq = 0.10 ± 0.05 at T ≤ 1050°C for the silicic sample plagioclase and based on the the distribution coefficient K_D(Fe-Mg)cpx-liq = 0.28 ± 0.08 for the clinopyroxene (Putirka 2008). We estimated the storage conditions of the magma based on phenocryst core and mantle analyses in equilibrium with the host rock. The corresponding depths have been derived assuming a 270 bar/km geobaric gradient (Best 2013).

The clinopyroxene in equilibrium with the US high-MgO samples host rock indicate mainly 3-5 kbar (mean 3.8 ± 1.4 kbar) and 1145-1202 °C (mean 1174 ± 45 °C) (Fig. 10A). The plagioclase in equilibrium with the US high-MgO samples host rock indicate 1.8-3.4 kbar (mean 2.8 ± 3.8 kbar) with a temperature of 1160-1170C° (mean 1163 ± 36 °C) (Fig. 10B). For the US low-MgO samples, the clinopyroxene in equilibrium with the host rock indicate 1.1-3 kbar (mean 2.1 ± 1.4 kbar) and 1126-1144 °C (mean 1132 ± 45 °C) (Fig. 10A). Regarding the US low-MgO samples plagioclase, the high An cores are in disequilibrium with their host rocks, while the lower An core and mantle (An66-82) are in equilibrium with the whole rock at 3.6-5.2 kbar (mean 4.3 ± 3.8 kbar) and temperatures of 1173-1184 °C (mean 1179 ± 36 °C) (Fig. 10B). Moreover, according to the KD equilibrium test, few plagioclase of one low-MgO sample with low An (An69-76) core analyses crystallized at very shallow depth (equilibrium at <1 km of depth, Fig. 10B). For the US rhyolitic sample, the plagioclase in equilibrium with the host rock indicate 1.1-4.6 kbar (mean 2.84 ± 3.8 kbar) and 961-977 °C (mean 968 ± 45 °C) (Fig. 10B).

No clinopyroxene of the CAG high-MgO samples are in equilibrium with the host rock. The plagioclase of the CAG high-MgO sample, on the contrary, in equilibrium with the host rock indicate 3.7-4.4 kbar (mean 3.9 ± 3.8 kbar) and a temperature of 1158-1164 °C (mean 1161 ± 36 °C) (Fig. 10B). The clinopyroxene in equilibrium with the CAG low-MgO samples host rock indicate mainly 3-5 kbar (mean 3.7 ± 1.4 kbar) and 1146-1175 °C (mean 1164 ± 45 °C) (Fig. 10A). The plagioclase in equilibrium with the low-MgO CAG samples host rock indicate 1.2-4 kbar (mean 2.2 ± 3.8 kbar) and a temperature of 1164-1178 °C (mean 1173 ± 36 °C) (Fig. 10B).

Chemical diffusion and crystallization kinetics in plagioclase

To the purpose of discussing mineral chemical zoning (chapter 5.2), it has to be taken into account that chemical diffusion and crystallization kinetics can affect the redistribution of the element from the melt to minerals. Mg diffusion is fast in plagioclase and can therefore obliterate the information about magma composition. However, considering that Mg diffuses faster than other elements treated in this work (i.e., CaAl–NaSi interdiffusion, Ti, Sr and Ba; Grove et al. 1984; Cherniak and Watson 2020; LaTourrette and Wasserburg 1998; Cherniak 2002; Giletti and Casserly 1994), the frequent variations observed for the MgO, as well as TiO2, Sr and Ba (Fig. 5, Fig. 7A; Fig. S3, Fig. S4), suggest that diffusion did not strongly affect the composition of the analyzed crystals. Kinetic effect, due to rapid growth of the crystals, could also affect the plagioclase mineral chemistry since the crystals do not have time to equilibrate with the hosting melt. This effect can be detected comparing the behaviour of compatible and incompatible elements (Ginibre et al. 2002; Bezard et al. 2017). Incompatible elements (Fe, Ba, Mg, Ti) will be enriched by the kinetic effect related to rapid growth, while compatible (Sr) will be depleted (Ginibre et al. 2002; Bottinga et al. 1966). In the analyzed US and CAG samples plagioclase, Sr and Ba behave coherently (Fig. 7A), increasing in the mantle/rim and arguing against the effect of kinetic processes. Therefore we interpret the variation in An, TiO2, Sr and Ba as related to changes in the magma composition.

Mg partition coefficient in plagioclase

In the case of Mg, recent work indicates that the Mg partitioning coefficient in plagioclase depends on the C1–I1 structural phase transition (An60 at 1000 C; Mutch et al. 2022). Mutch et al. (2022) observed a positive correlation in the C1 (albite) domain and a negative correlation in the I1 (anorthite) domain between An and MgO. Accordingly, the US and CAG plagioclase composition in the MgO vs An binary plot (Fig. S8) shows an increase in MgO before its concentration starts to decrease at about An65. The plagioclase chemical profiles confirm this behavior, with MgO decreasing when An decreases below An65 while increasing when above An65 (Fig. S3B). The observed MgO variations in the plagioclase seem thus to be related to the Mg mineral-melt partitioning and not to MgO concentration in the magma, confirming the finding of Mutch et al. (2022). For this reason, MgO variations have not been used for the interpretation.

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