Anatexis in K-poor and Si-Rich migmatites

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K-poor and Si-rich pelites are potential protoliths of migmatites, but often lack clear mineralogical indicators for anatexis. This is due to the limited number of peritectic phases, and because the products of melting reactions tend to be erased by retrograde deformation and recrystallization. Thus the pre- and post-peak assemblages may be almost identical and evidence for migmatization is overlooked.

The Roded migmatites of southern Israel were previously thought to form by subsolidus metamorphic differentiation based on the lack of K-feldspar in leucosomes and similarity of plagioclase compositions in coupled leucosomes and melanosomes [1]. Geochemical analysis of carefully sampled palaeosomes shows that the protoliths of Roded migmatites were K-poor and Si-rich metapelites. Nonetheless relict textural evidence for melting is found in the Roded migmatites including quartz-filled embayments in plagioclase crystals, and newly formed euhedral crystals of plagioclase in leucosomes. Likewise lenticular K-felsdpar occurs within melanosome biotite indicating muscovite dehydration melting. Thermodynamic analysis of specific palaeosome compositions show that anatexis should have occurred at peak P-T conditions estimated for the Roded migmatites (Fig.1; 4.5 kbar; ~650ºC). Petrographic and microstructural observations suggest that potassium was mobilized from the leucosomes to the melanosomes during melt crystallization: Leucosome K-feldspar was replaced by myrmekite and melanosome sillimanite was replaced by symplectites of muscovite and quartz.

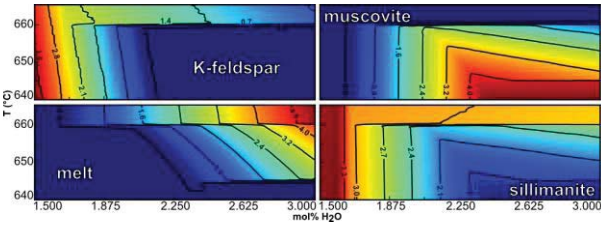

Figure 1: Modelled modes of key phases in melting reactions displayed on T vs. whole-rock H_2O content diagrams calculated for a specific rock composition of Roded migmatites. P=4.5 kbar.

Wider neosomes are characterized by diatexitic microstructures and by lower amount of myrmekite suggesting channeled influx of water resulting in higher degree of anatexis and dilution of potassium in the melt.

K-feldspar and muscovite+quartz symplectites in melanosomes, and myrmekite in leucosomes are established as criteria for inferring partial melting in heavily overprinted and retrogressively deformed K-poor, Si-rich migmatites.

[1] Gutkin & Eyal (1998) *Isr. J. Earth Sci.* **47**, 1-17.

Cation substitution in Iron Oxides in BIF-hosted Iron Ore: the Role of Inheritance and Fluids

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Little is known about the chemisty of the often complex iron oxides paragenetic sequences in BIF-hosted high-grade iron ore, and its potential use for genetic modelling (e.g., iron oxide and fluid origin) and exploration (e.g., as chemical tracer towards ore). In order to investigate this, minor and trace elements (lithophile: Ti, V, Cr, Sc, Al, Ga, Mg, Mn, and chalcophile: Ni, Cu, Zn, As, Co) in magnetite, hematite, and goethite species in the Koolyanobbing and Windarling iron ore deposits located in the Yilgarn Craton, which are hosted in lithostratigraphic unit of same Age (3.0 Ga), but in different greenstone belts, were analysed by means of laser ablation ICPMS.

Early very-low to low metamorphic grade magnetite (M-1) in least-altered BIF have lithophile element contents between 1 and 1000 ppm (Mg>Al>Mn>Ti>V>Ga) and chalcophile element contents between 0.1 and 100 ppm (Zn≥Ni≥Cu>Co±As). Compared to other hydrothermal deposit classes, magnetite types in BIF-hosted iron ore are poor in most substituting cations, especially Cr, V, and chalcophile elements (up to three magnitudes). The analytical precision decreases with lower element concentration (<100 ppm), but due to these very low concentrations, laser ablation is the only method to measure indicative chemistry of BIF-related iron oxides.

Hydrothermal magnetite (M-2) types in hydrothermal carbonate- and quartz-altered BIF (both are most prolific protoliths to ore) show typically even lower concentrations of most lithophile elements (Mg, Al, Ga, ±V, ±Ti) than M-1, but higher concentrations of Mn and chalcophile elements, most prominently Co. This suggests that in both deposits M-2 precipitated from fluids of a common type, probably low T (limited element substitution), sulphide-rich, and with a redox condition in which Mn substitution was favored. A partial inheritance of M-2 from disolved M-1 may be concluded from similar immobile element (V, Ti, Al, Ga) ratios.

Both, hydrothermal and supergene-related martite (after M-1 and M-2) do not involve significant changes of the chemistry, except a Mg (±Zn) loss (due to the exclusion of solely II-valent elements from hematite). A mineral chemical differentiation between hypogene and supergene martite is presently difficult, mainly due to the inheritance.

Intermediate to late stage hydrothermal hematite types show commonly a decrease of Mg, Mn, Zn, Ni, Cu, Co. In Koolyanobbing a remarkably similar content of other lithophile elements with respect to M-1 and/or M-2 points to a chemistry inheritance from (proximal) dissolved or in-situ replaced BIF magnetite.

Supergene goethite chemistry is distinct when comparing different goethite types (goethite after siderite vs quartz). For example, goethite after siderite shows similar Ni, Cu, and Co concentrations when compared to the replaced siderite.

In conclusion, cation substitution of hydrothermal and supergene iron oxides in BIF-related iron ore systems of the Yilgarn craton show that iron oxides are capable of capturing chemical signatures that help deciphering geological/fluid regimes.