The Formation of the Fe skarn deposit between Camibogazi and Arnastal (Gumushane, NE Turkey): Evidence from mineral chemistry

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The Fe skarn deposit in the Camibogazi and Arnastal area is located about 50 km north of Gumushane in NE Turkey. The deposit is hosted by Late Cretaceous Carbonate rocks which comprise dolomitic and sandy limestone, and limestone, adjacent to the Eocene Zigana Granitoid [1]. Petrographically, granitoid consists of syenogranite, monzogranite, quartz-monzonite and granite. Fe skarn deposit has mineral paragenesis of vesuvianite (idocrase), phlogopite, diopside, andradite, actinolite, tremolite, epidote, quartz, magnetite, hematite and less pyrrhotine and pyrite. Vesuvianite, diopside and phlogopite mineralization are only seen in a very narrow field of the Camibogazi Plateau (altitude of c. a. 2450 m). Magnetite, hematite, andradite, actinolite, tremolite, epidote, quartz and less parroting and pyrite are seen in Aransas Plateau.

Microprobe analyses indicate that majority of the skarn minerals are calcic and have high Mg/(Mg+Fe). Clinopyroxene consists of diopside ($Di_{96.66\text{-}99.04}$ $Hd_{0.67\text{-}2.80}$ $Jo_{0.29\text{-}0.55}$). Vesuvianite is mostly Mg-rich, and has very low MnO content (< 0.1 %). Phlogopite has high and nearly constant Mg/ (Mg+Fe) ratios, 0.93 to 0.94. The Ba content of phlogopite is low. Magnetite is generally massive, and some of them are martitization. According to vesuvianite-diopside geothermometry, the formation temperatures of skarn minerals in Camibogazi range from 329 to 558 ^{o}C.

The Fe exoskarn skarn deposit was characterized by early, high-temperature mineral assemblages dominated by anhydrous minerals diopside and garnet, and late low-temperature assemblages with hydrous minerals vesuvianite, phlogopite and epidote. Types of clinopyroxene and garnet, the abundance of magnetite with minor pyrite, and widespread occurrence of epidote suggest that the Fe skarn deposit between Camibogazi and Arnastal Plateau was formed under relatively oxidized conditions.

Reference

[1] Sipahi, F. (1996) MSc. Thesis, KTÜ FBE, Turkey (Unpub.).

Modelling pore water B and δ^{11}B signatures in the shallow subduction zone forearc: Examples from Barbados, Costa Rica and N. Japan

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At many subduction zones, pore water geochemistry from boreholes and mud volcanoes indicate a contribution from deep sources. These observations include pore water freshening, but enrichment in volatiles such as B and Li, and decreased δ^{11}B, δ^{6}Li, and δ^{37}Cl. Identifying the location and distribution of source regions for these tracers is one critical step toward characterizing subduction zone fluid transport. Here, we present a simple model to combine (1) heating and compaction that accompany progressive burial of sediment in subduction zones with (2) previously published laboratory experimental data that constrain the distribution coefficient (K_d) for B in marine sediments as a function of temperature, to quantify the expected distribution of [B] δ^{11}B within bulk mudstones in subduction zones. In a generic subduction zone with a taper angle of 8^{o}, we evaluate two end-member cases: a "cold" (60 mW m^{-2}) and a "warm" (120 mW m^{-2}) scenario. For the first, simulated values of [B] within the subducted sediment at 30 km from the trench range from 650–1200µM and values of δ^{11}B range from 23.4-30.6‰. At 60 km from the trench, [B] ranges from 1010-3340µM and δ^{11}B from 18.0-25‰. For the warm scenario, simulated [B] is 970-2400µM at 30 km and 2250-11480µM at 60 km; δ^{11}B ranges from 19.2–25.2‰ at 30 km and 15.92-19.5‰ at 60 km. These signatures are generally stronger than those observed in pore fluids at shallow depths, as would be expected due to probable re-adsorption of some B during updip or vertical advection. Ultimately, our results can be used for hydrologic models to quantitatively assess the fluid flow rates and permeabilities required to transport B at rates high enough to produce the observed signature.