

The role of bacteria in hydrogeochemistry, metal cycling and ore deposit formation

JAMES A. SAUNDERS

Dept. of Geology, Auburn University, Auburn, AL USA
36849 saundja@auburn.edu

Bacteria have both direct and indirect effects on the geochemical cycling of metal(loid)s. They may uptake metals into their cell walls or interiors and thus become colloid-size transporters of metals themselves. But most importantly, they facilitate electron transfers in elements that can have multiple oxidation states, such as metals, sulfur, and carbon. It is this capability that leads to formation of new minerals (some having the capacity to strongly sorb other metals), dissolution of existing ones, or causing the formation of aqueous complexes that can enhance the solubility of some metals. In aqueous environments, bacteria often control the ambient Eh and pH of the local setting as it moves toward transient (local) equilibrium conditions. Thus bacterial metabolism largely controls the major and trace element geochemical behavior in shallow groundwater settings and at least some low-temperature ore-forming environments. Recent research on bioremediation of metals provides additional support to the concept that bacteria can operate over timeframes and scales consistent with producing some types of metallic ore deposits. We have stimulated sulfate-reducing bacteria (SRB) in metal-contaminated groundwater to make relative insoluble Me-sulfide minerals that essentially remove virtually all of the metal contaminants. In the process, these nano-scale "bio-sulfide minerals" appear to have a high capacity to absorb metals even *after* formation. Textures of sulfide minerals formed by SRB during bioremediation (most notably pyrite and sphalerite) have textures reminiscent of those in certain sediment-hosted ores, supporting the concept along $\delta^{34}\text{S}$ values) that SRB may have been directly involved in forming ore minerals.

Iron isotope fractionation between hematite and aqueous fluids: Insights from hydrothermal experiments

G. SAUNIER, G.S. POKROVSKI AND F. POITRASSON

Université de Toulouse, CNRS, LMTG, 14 Av. E. Belin,
31400 Toulouse, France (saunier@lmtg.obs-mip.fr)

It is now possible to measure precisely natural Fe isotope variations. However, the first studies have shown the need for experimental calibrations to interpret these isotope data properly. For example, iron isotope analyses of hematite and pyrite from five hydrothermal ore deposits formed at 300-400°C show no Fe isotope fractionation between these two minerals ($\Delta^{57}\text{Fe}_{\text{pyrite-hematite}} = 0\%$). This is in marked contrast with theoretical predictions that yield equilibrium $\Delta^{57}\text{Fe}_{\text{pyr-hem}} = \sim 1\%$ at these temperatures [1]. Similarly, large discrepancies (up to 4‰ in $\Delta^{57}\text{Fe}$) are observed between experimental and theoretical equilibrium for Fe isotope fractionation between hematite and Fe^{3+} aqueous solution [2].

In an attempt to resolve this discrepancy, we have undertaken an experimental study of Fe isotope fractionation between hematite and a hydrothermal fluid. Our experiments consist in precipitating or dissolving hematite in a hydrothermal reactor in $\text{FeCl}_2\text{-NaCl-HCl-NaClO}_4$ aqueous solutions of known Fe isotope compositions. We monitor total Fe concentration and isotope ratio vs. time and temperature in the solution via periodically sampling.

In runs of hematite precipitation at 300°C and P_{sat} , a steady state for both total Fe concentration and $^{57}\text{Fe}/^{54}\text{Fe}$ ratio was attained after a few days. We observe an enrichment in the light isotope in the fluid relative to the mineral, with $\Delta^{57}\text{Fe}_{\text{fluid-hematite}} = \delta^{57}\text{Fe}_{\text{fluid}} - \delta^{57}\text{Fe}_{\text{hematite}} = -0.62 \pm 0.16\%$ (2SE). Dissolution runs at similar conditions yield $\Delta^{57}\text{Fe}_{\text{fluid-hem}} = -0.51 \pm 0.15\%$ (2SE). The agreement between these two sets of experiments suggests that Fe isotope equilibrium is likely to be attained. Our values are consistent with those calculated using the distribution of Fe aqueous species in the solution, dominated by Fe^{III} and Fe^{II} chloride complexes, and their theoretical equilibrium fractionation factors [1,3], $\Delta^{57}\text{Fe}_{\text{Fe(aq)-hematite}} = -0.64\%$. Further experiments are in progress on hematite and pyrite as function of temperature and solution composition to establish isotope fractionation laws between these minerals and hydrothermal fluids.

[1] Polyakov *et al.* (2007) GCA **71**, 3833-3846. [2] Skulan *et al.* (2002) GCA **66**, 2995-3015. [3] Schauble *et al.* (2001) GCA **65**, 2487-2497.