

## Experimental determination of Fe isotope fractionation between liquid metal, silicate and sulfide at high pressures and temperatures

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There is evidence for significant equilibrium Fe isotope fractionation ( $\sim 0.26$  ‰/amu) between metal and troilite (FeS) in iron meteorites and metal and olivine in pallasites [1-3]. However, it is not clear if the observed fractionations can be extrapolated to the pressure and temperature conditions of planetary core formation. We have investigated Fe isotope fractionation between silicate melt and liquid Fe-S alloys and between liquid iron and basaltic melt at pressure and temperature conditions of 2-2.5 GPa and 1650-2023 K using piston-cylinder partitioning experiments from previous studies [4-6]. Metal, sulfide and silicate fractions were separated from mounted and sectioned experimental charges by using a Micromill (New Wave). Sample dissolution, Fe purification and isotopic analysis followed established procedures [1]. In agreement with another experimental study [7] we find no appreciable fractionation between liquid iron metal and basaltic melt. However, there is a resolvable Fe isotope fractionation between silicate melt and Fe-S alloy which ranges from  $0.12 \pm 0.04$  to  $0.15 \pm 0.04$  ‰/amu for separate experiments. No relationship between apparent fractionation factor and pressure or temperature is detectable within the range covered by the experiments. The fractionation factors determined from our experiments overlap with the average equilibrium fractionation factor obtained between silicate melt and pyrrhotite [8] and are also broadly consistent with data from iron meteorites and pallasites. These observations suggest that resolvable stable isotope fractionation between Fe-S alloys and silicate melts can take place at extreme pressure and temperature conditions and that isotopically light Fe can be sequestered into the S-bearing parts of planetary cores.

[1] Williams *et al.* (2006) *EPSL* **250**, 486-500. [2] Zhu *et al.* (2002) *EPSL* **200**, 47-62. [3] Weyer *et al.* (2005) *EPSL* **240**, 251-264. [4] Kilburn & Wood (1997) *EPSL* **152**, 139-148. [5] Gessmann & Wood (2002) *EPSL* **200**, 63-78. [6] Wood *et al.* (in press) *EPSL*. [7] Poitrasson & Roskosz (2007) *LPSC XXXVIII*. [8] Schuessler *et al.* (2007) *GCA* **71**, 417-433.

## Applications of SHRIMP II to O isotopic analysis of granite zircon and conodont apatite

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Upgrades of the SHRIMP II ion microprobe (Cs primary ion source, oblique incidence electron gun, multiple collector) have given it the capability to measure light element isotopes (e.g. O, S) in electrically insulating minerals in negative ion mode. High count rates obtained with a 25  $\mu\text{m}$  diameter Cs<sup>+</sup> primary ion beam give analytical precisions of better than 0.1‰ per spot analysis, but do not alone guarantee high accuracy. That's required other modifications (sample mount redesign, nulling the Earth's magnetic field near the sample, analytical procedures for stabilising isotopic fractionation). The standard deviation of multiple analyses of isotopically uniform zircon or apatite is now routinely  $\sim 0.3$ ‰  $\delta^{18}\text{O}$ .

The Bega Batholith in SE Australia consists of  $\sim 130$  I- and A-type plutons exposed over  $\sim 9000$  km<sup>2</sup>. Systematic changes in the chemical and isotopic composition across the batholith (E-W) have variously been interpreted as reflecting changes in source rock composition or the mixing of different magmas as the axis of magmatism migrated eastward. Zircon U-Pb dating and O isotopic analysis of 30 plutons confirms a general eastward younging, but the range in  $\delta^{18}\text{O}$  (<3‰) is much smaller than expected from previous whole rock O analyses (>5‰) and there is no simple E-W trend.

The Ordovician saw the greatest evolutionary radiation of life in Earth's history, but the trigger remains unknown. A possible explanation is climate change. Previous studies of carbonate  $\delta^{18}\text{O}$  have estimated fluctuating Ordovician tropical sea surface temperatures of 50–70°C [1]. Our analyses of early Paleozoic bioapatite (conodont)  $\delta^{18}\text{O}$  suggest much lower temperatures ( $\sim 40$ – $25$ °C), with slow Early Ordovician cooling followed by a period of stability at modern equatorial temperatures, before rapid Late Ordovician cooling, possibly explaining the Middle Ordovician biodiversification and Late Ordovician extinctions. Temperatures returned to modern equatorial values in the Silurian.

[1] Veizer *et al.* (1999) *Chem. Geol.* **161**, 59-88.