#### FE AND MG SIGNATURES OF THE BONG URANIUM DEPOSIT, THELON BASIN, NUNAVUT

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## Abstract

The Bong deposit is one of several uranium occurrences located ca. 5 km southeast of the faulted margin of the Proterozoic Thelon sub-basin, within altered metasedimentary rocks of the Woodburn Lake group. The deposit exhibits features in common with unconformity-related deposits of the Athabasca Basin, in particular a spatial association with the regional unconformity, alteration and ore mineralogy and structural context. Primary uranium enrichment is associated with a broad hydrothermal alteration halo characterized by an inner illite-rich zone and an outer chlorite-rich zone. Whole rock and clay-sized fraction  $\delta^{25}$  Fe<sub>IRMM-014</sub> and  $\delta^{26}$  Mg<sub>DSM-3</sub> values of the least-altered host rocks cluster near 0 ‰, within range of average values of crustal materials. Zones characterized by intense illite alteration have high  $\delta^{57}$ Fe and  $\delta^{26}$ Mg values, with average whole-rock values of  $\delta^{57}$ Fe = 0.87 ±0.23‰ and  $\delta^{26}$ Mg = 0.61 ±0.29‰, and clay-size fraction values of  $\delta^{57}$ Fe = 0.77 ±0.29‰ and  $\delta^{26}$ Mg = 0.82 ±0.09‰. These isotopic values correlate with reductions in Fe<sup>2+</sup> concentrations (average of 0.47 mol. % versus average of 2.30 mol. %  $Fe^{2+}$  in least-altered hosts). The high  $\delta^{57}Fe$  values support leaching of  $Fe^{2+}_{(aq)}$  with low  $\delta^{57}Fe$  compositions from the system, causing precipitation of iron-bearing minerals with  $\delta^{s_7}$ Fe values higher than the host rocks. The  $\delta^{s_7}$ Fe values may have been locally influenced by the formation of hematite and Fe-oxyhydroxide minerals in secondary miniature roll-front mineralization due to redox reactions. The high 526 Mg values coincide with formation of illite, which preferentially incorporated heavier magnesium isotopes during the intense hydrothermal alteration observed in both the ore zone and an upper hematitic horizon. Despite the significant shifts in isotopic compositions, the net iron depletion in the ore zone via selective leaching of  $Fe^{2+}$  without  $Fe^{3+}$ enrichment implies that redox reactions were not the dominant mechanism of primary uranium mineralization at the Bong deposit. Mineralogical and geochemical data support primary uranium precipitation through reduction and neutralization of an oxidized, acidic fluid. As Fe<sup>2+</sup> was leached from the system in solution, enrichment of Fe-oxides and low whole-rock  $\delta^{57}$ Fe values may be a distal signature of these systems.

#### Introduction

Significant concentrations of hydrothermal Fe- and Mgbearing minerals (e.g. Fe-oxides/hydroxides and chlorite) are often associated with Proterozoic unconformity-related U systems, with their presence typically cited as a product of U precipitation mechanisms. For example, both basementand sandstone-hosted unconformity-related U deposits in the Athabasca Basin commonly exhibit an Mg-rich chlorite alteration halo and association with hematite (c.f. Hoeve and Sibbald, 1978; Quirt and Wasyliuk, 1997; Jefferson et al., 2007). One proposed mechanism for formation of Mg-rich chlorite (sudoite) associated with the deposits is by interaction between metamorphic biotite, amphibole and Fe-bearing chlorite present in the basement rocks with oxidizing, acidic fluids (e.g. Wallis et al., 1985; Alexandre et al., 2005). The Fe<sup>2+</sup> oxidized during the mineral transformations releases electrons that reduce U6+ in the oxidized fluids to immobile U<sup>4+</sup> (uraninite), while the liberated Fe<sup>3+</sup> and elements in the fluids form Fe-oxide minerals and sudoite.

The average isotopic composition of iron ( $\delta^{56}$ Fe<sub>IRMMS-014</sub> value) in the crust is 0.00 ± 0.08 ‰, which extends to most siliclastic sedimentary rocks plus alteration and weathering products (Heimann et al., 2008; Johnson et al., 2008; Craddock and Dauphas, 2010). However, large Fe fractionation can occur during redox reactions, changes in bonding states or leaching of Fe<sup>2+</sup> species under acidic conditions (Johnson et al., 2008). While ubiquitous Fe-oxyhydroxide minerals associated with basin-related U systems may reflect modern processes such as Fe-cycling during incursion of oxidized

waters, the structure, morphology and isotopic composition can be applied in combination to determine the conditions of formation (Diderikson et al., 2010).

Similar to Fe isotopes, the upper continental crust and mantle derived minerals have Mg isotopic compositions ( $\delta^{26}Mg_{DSM-3}$  values) between - 0.1 and - 0.3 ‰, with a mineralogical control on the fractionation of Mg between carbonates, clays and waters during low-temperatures processes such as weathering (Young and Galy, 2004; Pogge von Strandmann et al., 2008; Handler et al., 2009; Bourdon et al., 2010; Li et al., 2010; Wimpenny et al., 2014).

The hypothesis explored in this study is that by using mineralogical (primary minerals and alteration assemblages), textural (crystal size), Fe-speciation ( $Fe^{2+}$ ,  $Fe^{3+}$  molecular concentrations), whole-rock geochemistry and isotopic data (Mg and Fe isotopes) in concert, there is potential to define isotopic compositions that reflect redox reactions and changes in Fe and Mg bonding states characteristic of fertile hydrothermal U systems.

#### **Bong Deposit**

The Bong deposit is one of several U occurrences (Kiggavik deposit area) located ca. 5 km SE of the faulted margin of the Proterozoic Thelon sub-basin, within altered rocks informally termed the Woodburn Lake group (WLG; Fig. 1). Regionally, the WLG consists of highly deformed Neoarchean metasedimentary, metavolcanic and volcaniclastic strata, ca. 2.6 Ga mylonitized rhyolite (quartz eye rhylolite) and hybrid Lone Gull granite intrusions (1.83 –

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FIGURE 1. Regional geology of the Southeast Thelon Basin - Kiggavik deposit area, after Jefferson et al., (in prep.)

1.75 Ga; Scott et al. 2011) that are intruded by lamprophyre and syenitic dykes (Sharpe, 2013). At the deposit-scale, the host rocks are dominated by metagreywacke rocks interbedded with pelitic layers of the ca. 2711 Ma Pipedream assemblage, structurally intercalated with quartzite slices of early Paleoproterozoic Ketyet River group (Zaleski et al., 2000; Pehrsson et al., 2010; Sharpe, 2013; Scott et al., submitted; Fig. 2). All the deposits and showings in the Kiggavik area are located along approximately E-W trending faults that exhibit intense hydrothermal alteration (Fig.1). At the Bong deposit, the host metasedimentary rocks are fine- to medium-grained with subhorizontal foliation (Fig. 3). The primary minerals are quartz (60-70 modal %) and feldspar (15-25 modal %) with minor amounts of muscovite and biotite (ca. 5 modal %). The dominant feldspar is K-feldspar with minor plagioclase. Locally, there are layers that contain euhedral garnet porphyroblasts along with tabular biotite formed during peak metamorphism. Retrograde metamorphism resulted in pseudomorphic alteration of biotite to chlorite and sericitic alteration of feldspars along grain boundaries and fractures (Sharpe, 2013).

Uranium enrichment is associated with a broad alteration

halo characterized by illite and lesser chlorite with minor pyrite, rutile and apatite, the details of which are outlined in Sharpe (2013) and Rielger et al. (2014). Syn-ore illite alteration is centered on fault zones, replaces muscovite, biotite, metamorphic chlorite, garnet and K-feldspar and is associated with corrosion of quartz grains (Figs. 2 and 3). Syn-ore chlorite occurs as laths surrounded by an illite-rich matrix. As outlined by Sharpe (2013), U is proposed to have precipitated in two generations at the Bong deposit: Stage 1 veins associated with illite plus chlorite alteration and organic matter; and stage 2 miniature roll-fronts. These are cut by postore carbonate (calcite and dolomite) and drusy quartz veins 1-5 mm and up to 2 cm wide, respectively. The twenty-four samples examined under this study were selected from exploration drill core (Bong-042) characterized by clay alteration from depths of 236-315m, U enrichment plus clay alteration from 315-442m and red, hematitic horizons from 0-9m and 26-43m (Fig. 2). A hematite-bearing fault zone was intersected from 220-240m. Assay highlights from the hole include: 1.0m of 0.966% U starting at 351.5m, 19.5m of 0.688% U starting at 391.5m and 8.5m of 1.101% U starting at 425m (Blain and Morrison, 2008).



**FIGURE 2.** East-west cross-section of the Bong deposit, illustrating sampling locations from DDH Bong-042 (black boxes) examined in this study. Modified from Blain and Morrison (2008).

#### Methods

Samples for geochemical analysis were collected at 20 m intervals from a diamond drill hole that intersected alteration and mineralization at depth (Fig. 2). Representative samples (n=93) of the host rocks, alteration facies and ore mineralogy were also collected for petrography, the details of which are outlined in Sharpe (2013). The geochemical samples were pulverized until 95% passed through a 105 $\mu$ m sieve. Aliquots of the powdered material were then prepared for: 1) whole-rock geochemistry; 2) ferrous iron analysis; and 3) isotopic analysis.

The whole-rock geochemistry aliquot was mixed with a flux of lithium metaborate and lithium tetraborate and fused in an induction furnace. The molten melt was dissolved in a 5% nitric acid solution containing an internal standard. The



**FIGURE 3.** Representative photographs and corresponding photomicrographs (cross-polarized light) of drill core samples from Bong-042 illustrating the progression from least-altered (A and B) to highly illite-altered (G and H) metasedimentary host rocks at the Bong deposit. Il = illite, Ms = muscovite, Qz = quartz (from Sharpe, 2013).

samples were analyzed for major oxides and selected trace elements on a combination simultaneous/sequential Thermo Jarrell-Ash ENVIRO II ICP or a Varian Vista 735 ICP (Act-Labs ICP-MS Package 4B).

Ferrous iron contents were determined on whole-rock sample split by a titration method modified from Wilson (1955; ActLabs Package 4F). During titration, FeO was determined using a cold acid digestion of ammonium metavanadate and hydrofluoric acid in an open system. Ferrous ammonium sulphate was added after digestion and potassium dichromate was the titrating agent. When titrating, the endpoint was determined by colour. The contribution of ferrous iron, as FeO, was then subtracted from the total iron value determined by ICP-MS and the balance computed as ferric iron. The detection limit for FeO by titration is 0.01 wt. % (Act Labs, 2012).

Prior to Fe and Mg isotopic analysis, clay-sized fractions were made by centrifuge separation under distilled water.

Lab controls performed using a Particle Size Analyzer on clay-sized material indicated that approximately 95% of the recovered material is <0.004 mm, and that 85-90% is < 0.002 mm (Girard et al., 2004). Both whole-rock and claysized fractions were then subjected to a near-total four acid decomposition (HNO<sub>3</sub>, HF, HCl and HClO<sub>4</sub>). The mineralogy of whole-rock and clay-size separates was determined by X-ray powder diffraction analysis (XRD) at the Geological Survey of Canada. Suspensions (in water) of the samples were pipetted onto glass slides and air-dried overnight to produce oriented mounts. X-ray patterns of the air-dried samples were recorded on a Bruker D8 Advance Powder Diffractometer equipped with a Lynx-Eye Detector, Co Ka radiation set at 40 kV and 40 mA. Initial identification of minerals was made using EVA (Bruker AXS Inc.) software with comparison to reference mineral patterns using Powder Diffraction Files (PDF) of the International Centre for Diffraction Data (ICDD) and other available databases. Quantitative analysis is carried out using TOPAS (Bruker AXS Inc.), a PC-based program that performs Rietveld refinement (RR) of XRD spectra.

For isotopic analyses, the chromatographic separation of Mg was done by cation exchange resin (using a protocol adapted from Wombacher et al., 2008) while Fe was separated and purified by anion exchange (adapted from Chapman et al., 2005). In both cases, an aliquot of the elemental fraction obtained was analysed by ICP emission spectrometry to verify the purity of the fractions to avoid complicating matrix effects during the determination of the isotopic compositions and to ensure the >98% recovery necessary to avoid any potential fractionation during the chromatographic element separation. Procedural blanks were also analysed to ensure they were below the accepted limit of 1:1000 blank to analyte ratio. If this ratio was exceeded the digestion and chromatography was redone for that batch of samples.

Magnesium and Fe isotope ratios on samples, blanks and reference materials were determined using a Nu Plasma double-focusing multi-collector ICP-MS. A desolvating nebulization system (CETAC Aridus II) was used for sample introduction, with a 'micro-mist' nebulizer operated at 100  $\mu$ L/min flow rate for Mg and 50 uL/min for Fe. The use of the Aridus II reduced the 12C14N interference on 26Mg to insignificant levels while the 12C dimer interference on 24Mg could also be kept to insignificant levels by cleaning the sampler cone daily. Background signals coming from <sup>40</sup>Ar<sup>14</sup>N, <sup>40</sup>Ar<sup>16</sup>O and <sup>40</sup>Ar<sup>16</sup>O<sup>1</sup>H on <sup>54</sup>Fe, <sup>56</sup>Fe and <sup>57</sup>Fe respectively were reduced to <0.03% ArN:54Fe, <0.05 ArO:56Fe and <0.005% ArOH:57Fe by using lower flow rates and very careful optimization of the Aridus II gas flows. A 10<sup>10</sup> ohms amplifier (instead of the typical 10<sup>11</sup> ohm amplifier) was used for the detection of 56Fe which changed the upper limit of the detector from 10 V to 100 V. This modification permitted the use of higher concentrations of Fe during multicollector analyses which augmented the background to signal ratio. The use of a desolvator coupled to a lower nebulizer flow rate and higher background to signal ratio yielded the precision required to perform isotopic analysis on Fe. For both Mg and Fe isotope analyses, correction for background was done by measuring an on-peak zero and monitored on a regular basis to ensure the correction was valid throughout individual runs. Instrumental drift was corrected by using a standard sample bracketing technique (SBB protocol). Magnesium isotopes were calibrated to the DMS3 standard while Fe isotopes to the IRMM014 standard. Concentrations of standards and samples were matched to generate detector signals with no more than 10% difference. Typical concentrations for Mg analyses were ca. 75 ppb, whereas Fe was ran at ca. 500 ppb to ensure signal strengths significantly elevated above background. Average two standard deviation errors indicated by long-term repeat analyses of the isotopic standards are  $\pm 0.06$  ‰ for  $\delta^{57}$ Fe,  $\pm 0.05$  for  $\delta^{56}$ Fe and  $\pm 0.07$  ‰ for  $\delta^{26}$ Mg.

## Results

#### Mineralogical controls

Illite from the ore-bearing alteration zones at the Bong deposit comprises more than 70 modal % of the rock. In these zones, illite contains on average 2.87 and 2.35 wt. % FeO and MgO, respectively, (Sharpe, 2013; Figs. 3G, H). In the illite zone, hematite can locally constitute 10 modal % of the rocks and is the dominant host of Fe. The illite zones contain trace to minor amounts of sudoite (chlorite) intergrown with illite, based on XRD of clay-size fractions in this study and previous work (Sharpe, 2013; Riegler et al., 2014). Just above the illite-bearing ore zone, the clay alteration assemblage is characterized by increased concentrations of sudoite (c.f. Riegler et al., 2014) and carbonate (calcite and dolomite) veins that are the dominant Mg-bearing phases.

## *Fe and Mg isotopic signatures*

Isotopic results are reported as  $\delta^{57}$ Fe and  $\delta^{26}$ Mg, in per mil (‰) relative to Isotopic Reference Material (IRMM-014; Taylor et al., 1992) and Dead Sea Magnesium (DSM-3; Galy et al., 2003):

$$\begin{split} \delta^{^{56,\,57}}Fe &= ({}^{^{56,\,57}}Fe/{}^{^{54}}Fe_{Sample} \ / \ {}^{^{56,\,57}}Fe/{}^{^{54}}Fe_{IRMM-014} - 1)*10^3 \\ \delta^{^{26}}Mg &= ({}^{^{26}}Mg/{}^{^{24}}Mg_{Sample} \ / \ {}^{^{26}}Mg/{}^{^{24}}Mg_{DSM-3} - 1)*10^3 \end{split}$$

The whole rock isotopic data reveal two zones with significantly higher Fe and Mg isotopic values: 9-29m and 319–439m (Fig. 4, Appendices I and II). Predictably, carbonate-bearing (calcite and dolomite) samples have much lower  $\delta^{57}$ Fe and  $\delta^{26}$ Mg values (259–299m and 339m). For both zones, average whole-rock values are:  $\delta^{57}Fe = 0.87$  $\pm 0.23\%$  and  $\delta^{26}Mg = 0.61 \pm 0.29\%$ ; Fig. 4). Averages of the less-altered (in terms of mineralogy and chemistry) wholerock data are:  $\delta^{57}$ Fe = 0.20 ±0.09‰ and  $\delta^{26}$ Mg = -0.19  $\pm 0.25\%$ . In the two zones with high values, average claysized fraction values are:  $\delta^{57}$ Fe = 0.77 ±0.29‰ and  $\delta^{26}$ Mg =  $0.82 \pm 0.09\%$ ; Fig. 4). The average less-altered, clay-sized fraction results are:  $\delta^{57}Fe = 0.35 \pm 0.27\%$  and  $\delta^{26}Mg = 0.06$  $\pm 0.21\%$ . Plotting of  $\delta^{57}$ Fe versus  $\delta^{56}$ Fe and  $\delta^{25}$ Mg versus  $\delta^{26}$ Mg values for both whole-rock and clay fractions yielded linear trends consistent with mass dependent isotopic fraction, with slopes of 1.4831 and 0.5228, respectively.

The clay-sized fraction produced more variable Fe isotopic results, including an additional zone of high  $\delta^{57}$ Fe values associated with the hematite-bearing fault sampled at



FIGURE 4. Depth profiles of (A)  $\delta^{57}$ Fe isotope results for whole-rock (red data points and lines) and clay-size fractions (black points and lines) and (B)  $\delta^{26}$ Mg results for both whole-rock (red points and lines) and clay-size fractions (black points and lines).

229m. Conversely, the Mg isotopic data from the clay-sized fractions has less spread in the data, with least altered rocks yielding  $\delta^{26}$ Mg values near 0 ‰ (Fig. 4B).

## Fe oxidation states

A plot of total Fe versus depth (Fig. 5A) indicates significant changes in the total Fe contents and oxidation states, with a reduction in total Fe (wt. %) in the upper red, hematitic and U-bearing zones at depths of 9m, 29, and 315–442m (Fig. 2). Figure 5B illustrates that this reduction in total Fe reflects significant reductions in Fe<sup>2+</sup> concentrations (0.47 vs. 2.30 mol. Fe<sup>2+</sup>) while Fe<sup>3+</sup> contents remain relatively constant.

## Discussion

# Interpretation of isotopic data

Whole rock and clay-sized fraction  $\delta^{57}$ Fe and  $\delta^{26}$ Mg values of the least altered rocks in the study cluster near 0 ‰, within range of average values of crustal materials (c.f. Young and Galy, 2004; Johnson et al., 2008; Craddock and Dauphas, 2010). High  $\delta^{26}$ Mg and  $\delta^{57}$ Fe values in both wholerock and clay-sized fractions are clearly associated the intense hydrothermal alteration at the Bong deposit that resulted in the breakdown of the pre-existing mineral assemblages (changes in bonding states for Fe and Mg) and significant leaching of Fe<sup>2+</sup>. However, when examined in detail, only the U-bearing, illite-altered samples and red, hematitic horizons have high  $\delta^{26}$ Mg and  $\delta^{57}$ Fe values — the clayaltered zone above the U-bearing illite alteration (236–315m) has lower values, reflecting either the presence carbonate minerals enriched in lighter Mg relative to their formational fluids or the presence of sudoite as noted by



**FIGURE 5.** Depth profile of (A) total iron (FeO +  $Fe_2O_3$ ) and (B) molecular proportions of  $Fe^{2*}$  and  $Fe^{3*}$ , after Sharpe (2013).

Riegler et al. (2014). Similar to carbonate minerals, chlorite has been shown to preferentially incorporate isotopically light Mg relative to the host rocks and fluids (Ryu et al., 2011). Leaching of Fe as  $Fe^{2+}$  species can produce net positive shifts in the whole-rock data as the relatively stable  $Fe^{3+}$ bearing minerals preferentially incorporate the heavier Fe isotopes (Johnson et al., 2008), while clay minerals formed during such processes (i.e. illite) preferential incorporate isotopically heavier Mg (Tipper et al., 2006, 2012; Opfergelt et al., 2012; Wimpenny et al., 2014). The positive shift in  $\delta^{57}Fe$ values in the ore zone may have been locally enhanced by the formation of hematite and Fe-oxyhydroxide minerals in secondary miniature roll-front mineralization due to redox reactions. The high  $\delta^{s7}$ Fe values observed in the clay-sized fraction from the hematite-bearing fault may record very low temperature processes such as modern Fe-silicate weathering by oxidized meteoric fluids along faults (c.f. Diderikson et al., 2010), but further studies are required to verify the Fe mineralogical controls on the isotopic composition of the clay-sized fraction.

## Linking isotopic data to processes

The upper hematitic horizon at the Bong deposit was originally thought to represent a paleoweathering horizon, much like the possibly coeval lateritic profile observed along the unconformity at the base of the Athabasca Group (Mac-Donald, 1980; Gall, 1994). Weathering under oxic conditions generates insoluble Fe<sup>3+</sup> secondary silicate and oxide minerals and can produce large increases in  $Fe^{3+}/Fe^{2+}$  ratios. However, the stability of Fe<sup>3+</sup>-bearing minerals typically results in little whole-rock Fe isotope fractionation because loss of mobile Fe can be insignificant (Beard et al. 2003; Kiczka et al., 2011). Therefore, the similarity in isotopic values and lack of enrichment in relatively stable Fe<sup>3+</sup>-bearing minerals in the clay-altered and hematitic zones suggests that the red, hematitic horizons near the present surface are unlikely to have formed through intense paleoweathering processes. Gall (1994) characterized the Thelon paleosol as having lost TiO<sub>2</sub>, total Fe, Fe<sup>2+</sup> and K<sub>2</sub>O, and enrichment in  $Fe^{3+}$ . In comparison to the least-altered units, the hematitic horizons at the Bong deposit are characterized by losses in total Fe (via leaching of Fe<sup>2+</sup>) and MgO, no gains in Fe<sup>3+</sup> and gains in K<sub>2</sub>O. The increase in K<sub>2</sub>O contents reflects the presence of illite, a feature ascribed to K-metasomatism by Matthews & Scharrer (1968), Erikson and Soegaard (1985) and Gall (1994) but herein proposed as a product of hydrothermal alteration. Furthermore, the uppermost hematitic zone has  $\delta^{26}$ Mg values of -0.20‰ (whole rock) and -0.09 ‰ (clay fraction) plus contains K-feldspar; features similar to the less-altered host rocks and inconsistent with a lateritic paleoweathering origin for the hematitic zones.

As noted by Sharpe (2013), the proposed fluid chemistry during the primary U event (pH<4.5 at 200°C; Fig. 6) would have to be suitable for leaching of Fe<sup>2+</sup><sub>(aq)</sub> with low  $\delta^{57}$ Fe compositions from the system, causing precipitation of alteration minerals with  $\delta^{57}$ Fe values higher than the host rocks. As depicted in Figure 6, the destruction of the host rock mineral assemblages (feldspars, muscovite, biotite, metamorphic chlorite) and leaching of Fe<sup>2+</sup> are consistent with the incursion of a warm (~225°C), oxidized, U-bearing, acidic fluid that was neutralized and reduced through interactions with the basement rocks, resulting in the formation of illite, uraninite, chlorite and pyrite. These processes are consistent with the shifts in mineralogy, geochemistry, and Fe and Mg isotopic data.

#### **Implications for Exploration**

The isotopic data clearly reflect fertile hydrothermal alteration in both the U-bearing zone and upper hematitic



**FIGURE 6.** Log  $fO_2$  - pH diagram illustrating the distribution and stability of U and Fe phases and K- and Mg-silicates in an aqueous system containing 10 ppm Fe, 100 ppm F, 1000 ppm S, 1000 ppm K, 100 ppm Mg and 1m NaCl at P<sub>CO3</sub> of 10 atm and 200°C. Sch = schoepite. Shaded region denotes UO<sub>2</sub> field, and arrow approximate fluid pathway for primary U deposition: (1) Fe<sup>2+</sup> leaching, and (2) formation of illite, reduction and neutralization of fluids resulting in uraninite and pyrite precipitation. Modified from Romberger (1984)

horizon. The net Fe depletion in the ore zone via selective leaching of Fe<sup>2+</sup> without Fe<sup>3+</sup> enrichment implies that redox reactions were not the dominant mechanism of primary U precipitation. As Fe<sup>2+</sup> was leached from the system in solution, enrichment of Fe-oxides may be a distal signature of these systems – a hypothesis supported by the presence of significant hematite alteration along key regional faults in the district (Fig. 1). The high  $\delta^{26}$ Mg and  $\delta^{57}$ Fe values, presence of illite, and leaching of Fe<sup>2+</sup> supports the hypothesis that the upper hematitic zone likely formed under conditions similar to the primary U event and thus may be considered as a potential indication of nearby U mineralization. Given the structural control on alteration and mineralization, expression of these signatures in fault systems may provide a means of evaluating prospective drill targets. On the deposit-scale, these results indicate that Fe and Mg isotopic data highlight fertile clay alteration and when coupled with geochemical and mineralogical data, provide constraints on the genesis of these U systems. In terms of signatures relevant to exploration, U-bearing zones and prospective fault systems are best characterized by an outer chlorite halo and an inner, highly altered illite zone that results in Fe depletion coupled with high  $\delta^{26}$ Mg and whole-rock  $\delta^{57}$ Fe values and nominal increases in the isotopic values in the clay-sized  $\delta^{57}$ Fe values (Fig. 7).



FIGURE 7. Alteration mineralogy at depth in relation to whole-rock  $\delta^{s_7}$ Fe (A) and clay-size fraction  $\delta^{26}$ Mg (B) signatures and changes in Fe oxidation states (C).

# **Future Work**

Results are pending from sequential leach experiments, using 1M NH<sub>4</sub>OAc, 1M NH<sub>2</sub>OH, and four acid solutions to evaluate isotopic contributions from carbonate minerals, adsorbed species and fine-grained Fe-oxyhydroxides. Analyses are also underway on samples from the McArthur River deposit, with linkages to a detailed isotopic-mineralogical study by Acevedo and Kyser (this volume).

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	Depth	δ⁵6Fe	$\pm 2\sigma$	δ <sup>57</sup> Fe	$\pm 2\sigma$	$\delta^{26}Mg$	$\pm 2\sigma$	δ²5Mg	$\pm 2\sigma$
sample ID	(m)	IRMM014		IRMM014		DSM3		DSM3	
Whole-rock									
RS-11-081	9	0.48	0.07	0.74	0.05	-0.20	0.10	-0.11	0.13
RS-11-083	29	0.60	0.07	0.92	0.11	0.69	0.16	0.36	0.16
RS-11-085	49	0.12	0.02	0.18	0.01	0.28	0.08	0.17	0.05
RS-11-087	69	0.16	0.05	0.23	0.01	-0.21	0.10	-0.08	0.13
RS-11-089	89	0.06	0.05	0.10	0.09	-0.24	0.01	-0.14	0.02
RS-11-091	109	0.11	0.03	0.20	0.03	-0.33	0.11	-0.18	0.05
RS-11-093	129	0.16	0.07	0.24	0.08	-0.42	0.09	-0.19	0.04
RS-11-095	149	0.10	0.10	0.16	0.09	-0.16	0.03	-0.09	0.02
RS-11-097	169	0.03	0.05	0.03	0.05	0.02	0.14	0.00	0.04
RS-11-099	189	0.15	0.05	0.22	0.08	-0.10	0.12	-0.07	0.06
RS-11-101	209	0.11	0.04	0.14	0.04	-0.16	0.23	-0.09	0.05
RS-11-103	229	0.26	0.05	0.39	0.06	0.19	0.19	0.12	0.22
RS-11-113	239	0.21	0.05	0.30	0.04	0.09	0.07	0.04	0.05
RS-11-115	259	0.08	0.04	0.15	0.06	-0.40	0.08	-0.21	0.03
RS-11-117	279	0.04	0.05	0.10	0.07	-0.57	0.10	-0.31	0.06
RS-11-119	299	0.10	0.03	0.15	0.08	-0.61	0.08	-0.32	0.07
RS-11-121	319	0.54	0.07	0.80	0.03	0.93	0.05	0.47	0.03
RS-11-123	339	-0.41	0.06	-0.59	0.08	0.14	0.13	0.05	0.06
RS-11-125	359	0.33	0.02	0.48	0.01	0.75	0.16	0.39	0.05
RS-11-127	379	0.61	0.05	0.89	0.06	0.45	0.12	0.25	0.06
RS-11-129	399	0.64	0.04	0.97	0.07	0.56	0.06	0.29	0.06
RS-11-131	419	0.59	0.06	0.89	0.08	1.02	0.19	0.54	0.12
RS-11-133	439	0.85	0.05	1.30	0.07	0.70	0.11	0.34	0.06
RS-11-135	459	0.26	0.03	0.39	0.05	0.18	0.09	0.07	0.02

Appendix I. Whole-rock isotopic data from drill hole Bong-042

Appendix II. Clay-size fraction isotopic data from drill hole Bong-042.

	Depth	δ <sup>56</sup> Fe	$\pm 2\sigma$	δ <sup>57</sup> Fe	$\pm 2\sigma$	δ <sup>26</sup> Mg	$\pm 2\sigma$	δ²5Mg	$\pm 2\sigma$
sample ID	(m)	IRMM014		IRMM014		DSM3		DSM3	
Clay-size frac	ction								
RS-11-081C	9	0.68	0.07	1.00	0.11	-0.09	0.10	-0.03	0.10
RS-11-083C	29	0.76	0.05	1.12	0.06	0.77	0.10	0.41	0.10
RS-11-085C	49	0.17	0.02	0.27	0.02	0.78	0.10	0.40	0.10
RS-11-087C	69	0.44	0.11	0.67	0.08	-0.07	0.10	-0.03	0.10
RS-11-089C	89	0.13	0.07	0.20	0.10	-0.01	0.10	-0.02	0.10
RS-11-091C	109	0.33	0.08	0.49	0.11	-0.25	0.10	-0.13	0.10
RS-11-093C	129	0.24	0.01	0.36	0.04	-0.19	0.10	-0.10	0.10
RS-11-095C	149	0.20	0.03	0.29	0.05	-0.12	0.10	-0.08	0.10
RS-11-097C	169	0.09	0.09	0.11	0.10	0.17	0.16	0.11	0.11
RS-11-099C	189	0.18	0.08	0.27	0.09	-0.04	0.10	-0.02	0.10
RS-11-101C	209	0.22	0.09	0.33	0.08	0.19	0.10	0.08	0.10
RS-11-103C	229	0.78	0.00	1.15	0.03	0.41	0.12	0.20	0.10
RS-11-113C	239	0.26	0.04	0.37	0.05	0.48	0.11	0.26	0.05
RS-11-115C	259	0.19	0.09	0.27	0.12	0.20	0.10	0.10	0.10
RS-11-117C	279	0.03	0.07	0.05	0.12	-0.03	0.11	-0.02	0.10
RS-11-119C	299	0.02	0.13	0.07	0.18	0.06	0.22	0.02	0.11
RS-11-121C	319	0.71	0.12	1.08	0.13	0.78	0.10	0.40	0.10
RS-11-123C	339	-0.05	0.19	-0.07	0.33	0.75	0.10	0.38	0.10
RS-11-125C	359	0.21	0.10	0.32	0.12	0.88	0.10	0.46	0.10
RS-11-127C	379	0.40	0.08	0.60	0.14	0.73	0.22	0.40	0.11
RS-11-129C	399	0.45	0.20	0.70	0.33	0.81	0.15	0.41	0.10
RS-11-131C	419	0.31	0.12	0.48	0.16	1.02	0.10	0.53	0.10
RS-11-133C	439	0.58	0.04	0.85	0.14	0.89	0.10	0.47	0.10
RS-11-135C	459	0.21	0.13	0.34	0.21	0.23	0.10	0.11	0.10