AUBRITES: AN IRON AND ZINC ISOTOPE STUDY

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Introduction: Aubrites are highly reduced meteorites, a property they share with the enstatite chondrites. Other similarities between the two groups are O-isotope signatures [1], low FeO content, mineralogy and mineral composition [2,3]. However, significant differences between the two groups point towards origin of the aubrites on a parent body distinct from that of the enstatite chondrites [4]. It has been suggested that aubrites formed by either direct condensation [5-7] or from partial melting of an enstatite-like precursor [3]. Evidence of an igneous origin includes large crystal sizes [up to 10mm, 3], melt inclusions in enstatite [8], europium anomaly [9] and mineralogy consistent with fractional crystallization from an Mg-rich magma [10,11]. The aubrite parent body may have been involved in a major collision, which resulted in disruption of the body, followed by gravitational reassembly into a rubble pile body [10].

Rationale: Iron is a moderately volatile element with a 50% condensation temperature of 1334K [12]. Zinc is more volatile than iron with a 50% condensation temperature of 726K [12]. Iron isotopes are minimally fractionated during igneous processing [13] and may therefore be used to characterise the nature of the protolithic material. Zinc is more volatile than iron and may be fractionated during impacting. Combining both of these isotope systems may help to constrain the history of the aubrite parent body.

Method: The method for analysing iron and zinc isotopes by multiple collector inductively coupled plasma mass spectrometry has been described previously [14,15]. The precision at the 2σ level achieved for iron is $\delta^{56}Fe \pm 0.06‰$ and $\delta^{57}Fe \pm 0.08‰$ and for zinc is $\delta^{66}Zn \pm 0.05‰$ and $\delta^{68}Zn \pm 0.07‰$.

Samples: The sample set includes Aubres, Bishopville, Bustee, Cumberland Falls, Khor Temiki, Mayo Belwa, Norton County, Peña Blanca Spring and Shallowater. Apart from Bustee and Khor Temiki, which are regolith breccias, all samples are fragmental breccias. Shallowater is believed to derive from a parent body separate from the remaining aubrites due to its highly individual characteristics [16-17] and its iron and zinc isotope composition is considered separately.
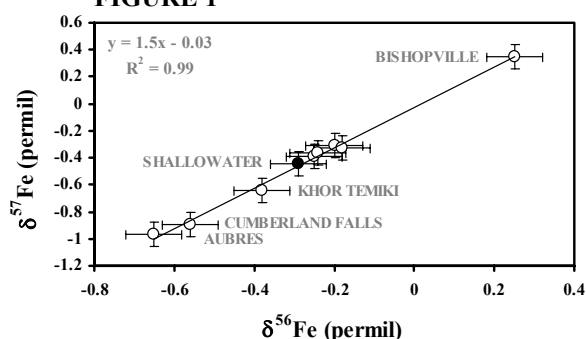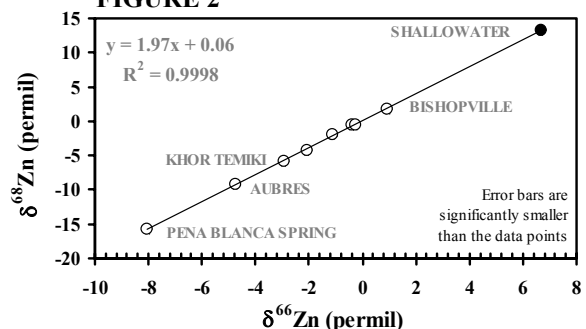
Iron Isotope Fractionation: The range of fractionation in aubrites ranges from 0.1 to 0.33 permil/amu (Figure 1). Apart from Bishopville ($\delta^{56}Fe = +0.25‰$), all samples are isotopically light ($\delta^{56}Fe = -0.65$ to $-0.18‰$). Four aubrites cluster between $\delta^{56}Fe$ = -0.25 to $-0.18‰$ (Norton County, Bustee, Mayo Belwa and Peña Blanca Spring). Aubres is the most isotopically light sample ($\delta^{56}Fe = -0.65‰$), followed by Cumberland Falls ($\delta^{56}Fe = -0.56‰$) and Khor Temiki ($\delta^{56}Fe = -0.38‰$). Shallowater, which represents a different parent body, is similarly fractionated to the cluster ($\delta^{56}Fe = -0.29‰$).

Zinc Isotope Fractionation: The range of fractionation of zinc isotopes is 0.13 to 4 permil/amu. Both Bishopville ($\delta^{66}Zn = +0.94‰$) and Shallowater ($\delta^{66}Zn = +6.68‰$) are isotopically heavy. The remaining samples are isotopically light ranging from $\delta^{66}Zn = -8.04‰$ (Peña Blanca Spring), $\delta^{66}Zn = -0.26‰$ (Cumberland Falls).

Discussion:

Zinc Isotopes: The significant range of fractionation observed may indicate that volatilisation of zinc has occurred at some point in its history. However, the fractionation systematics are not easily explained as many of the zinc isotope compositions are isotopically light. The opposite is expected if the signatures were generated by the impact event that resulted in collisional break-up of the aubrite parent body. It is possible that the initial isotope signatures were lighter than presently observed.

FIGURE 1

$y = 1.5x - 0.03$
$R^2 = 0.99$

FIGURE 2

$y = 1.97x + 0.06$
$R^2 = 0.9998$

Error bars are significantly smaller than the data points

Iron Isotopes: The isotopes of iron are not significantly fractionated during igneous processing of terrestrial igneous rocks [e.g. 13]. The range of iron isotope fractionation seen in aubrites may be explained in a number of ways. Firstly, the aubrites may originate on separate parent bodies and will thus represent different iron isotope reservoirs. However, both oxygen isotopes [1] and the unusual compositional characteristics of these meteorites suggests that, at the very least, they originated in the same region of the solar nebula. If the aubrites analysed here are from different parent bodies, then it is likely that these bodies are spatially related.

Secondly, it is possible the aubrites were formed on the same parent body but from a heterogeneous protolith and that igneous processing did not homogenise the iron isotope signature. If melting on the aubrite parent body occurred in isolated pockets and the melted protolith contained material with disparate iron isotope signatures then aubrites generated from localised melting could have preserved heterogeneous iron isotope signatures. However, in order to generate melts of aubritic composition and also to allow the bulk of the metal to segregate to the core, melting has to be extensive. This would suggest that the melt attained a homogenous iron isotope signature.

Finally, it is possible that an impact event may have resulted in fractionation of the zinc isotopes and contributed to fractionation of the iron isotopes. Except for three outliers (Shallowater, Cumberland Falls and Peña Blanca Spring), the fractionation systematics of iron and zinc isotopes are well correlated (Figure 3). Shallowater is atypical because it originated on a different parent body. The isotopic signature of Cumberland Falls may lie off the correlation line due to the presence of chondritic clasts, which were admixed during surface residence [18,19]. Peña Blanca Spring has a lighter zinc isotope composition with respect to the other aubrite samples and may have resided deeper within the parent body and consequently was shielded from the intensity of the impact event.

Fragmental Breccias vs. Regolith Breccias: Bustee and Khor Temiki are regolith breccias and the remaining samples (excluding Shallowater) are fragmental breccias. The iron and zinc isotope signatures of both regolith breccias are different from each other, but do not differ significantly from the isotope systematics of the fragmental breccias, indicating that residence on or near the surface of the parent body has not significantly altered the isotopic composition of iron or zinc.

FIGURE 3
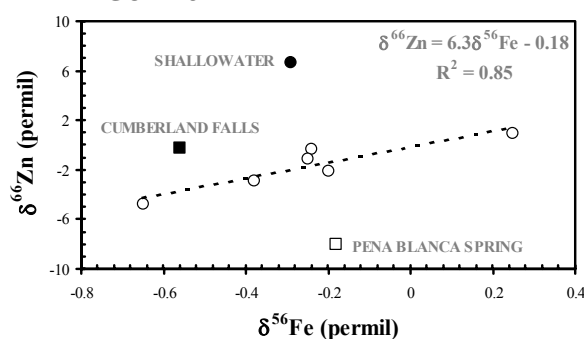
Bishopville: Excluding Shallowater data, Bishopville contains the most isotopically heavy iron and zinc in this sample set. One interpretation is that the isotopic composition of Bishopville was significantly modified an impact event, volatilising both the lighter isotopes of iron and zinc.

Shallowater: Shallowater experienced a very complex cooling history [16,17]. It is believed to have formed on a molten or partly molten, differentiated parent body which was disrupted by low velocity impact with a solid body. The impact-generated debris was subsequently reassembled into the Shallowater parent body [16,17]. Iron isotope compositions fall within the range observed for the other aubrites. Zinc isotopes are significantly isotopically heavy which indicates volatilisation of zinc during the formation of Shallowater. In addition to the impact history of the Shallowater parent body, Keil et al. [17] note that the Shallowater melt may have been superheated, to above the melting point of pure enstatite (1853K). If this caused overall loss of Zn, then the impact processing may have driven the isotopic composition of zinc towards such a heavy signature, but does not appear to have modified the iron isotope composition.

References: [01] Clayton et al. 1984, LPSC 15th C245-9. [02] Keil, 1968, JGR 73, 6945-76. [03] Watters and Prinz, 1979, LPSC 10th 1073-93. [04] Keil, 1989, Meteoritics 24, 195-208. [05] Wasson and Wai, 1970, GCA 3, 169-84. [06] Richter et al., 1979, LPSC 10th 1028-30. [07] Sears, 1980, Icarus 43, 184-202. [08] Fuchs, 1974, Meteoritics 9, 342. [09] Schmitt et al., 1964, GCA 28, 67-86. [10] Okada et al., 1988, Meteoritics 23, 59-74. [11] Taylor et al., 1988, LPSC 19th 1185-6. [12]Lodders, 2003, Astro. Jl. 591, 1220-1247. [13] Beard & Johnson, 1999, GCA 63, 1653-60. [14] Mullane et al., 2003, In: Plasma Source Mass Spectrometry, Royal Society of Chemistry, 351-61. [15] Mason et al., 2004, J. Anal. At. Spectrom. 19, 209-17 and 218-26. [16] Keil, 1988, Meteoritics 23, 278-9. [17] Keil et al., 1989, GCA 53, 3291-307. [18] Binns, 1969, Astrophys. Space Sci. Lib. 12, 696. [19] Kallemeyn & Wasson, 1985, GCA 49, 261-70.