Earth and Planetary Science Letters 482 (2018) 490–500

Contents lists available at ScienceDirect

Earth and Planetary Science Letters

www.elsevier.com/locate/epsl

Late metal–silicate separation on the IAB parent asteroid: Constraints from combined W and Pt isotopes and thermal modelling

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ARTICLE INFO

Article history:
Received 27 June 2017
Received in revised form 13 November 2017
Accepted 14 November 2017
Available online xxxx
Editor: F. Moynier

Keywords:
IAB iron meteorites
Hf–W chronometry
Pt isotopes
multi-collector ICP-MS
Planetesimals
thermal histories

ABSTRACT

The short-lived ^{182}Hf–^{182}W decay system is a powerful chronometer for constraining the timing of metal–silicate separation and core formation in planetesimals and planets. Neutron capture effects on W isotopes, however, significantly hamper the application of this tool. In order to correct for neutron capture effects, Pt isotopes have emerged as a reliable in-situ neutron dosimeter. This study applies this method to IAB iron meteorites, in order to constrain the timing of metal segregation on the IAB parent body.

The ε^{182}W values obtained for the IAB iron meteorites range from -3.61 ± 0.10 to -2.73 ± 0.09. Correlating ε^{i}Pt with ε^{182}W data yields a pre-neutron capture ε^{182}W of -2.90 ± 0.06. This corresponds to a metal–silicate separation age of 6.0 ± 0.8 Ma after CAI for the IAB parent body, and is interpreted to represent a body-wide melting event. Later, between 10 and 14 Ma after CAI, an impact led to a catastrophic break-up and subsequent reassembly of the parent body. Thermal models of the interior evolution that are consistent with these estimates suggest that the IAB parent body underwent metal–silicate separation as a result of internal heating by short-lived radionuclides and accreted at around 1.4 ± 0.1 Ma after CAIs with a radius of greater than 60 km.

1. Introduction

The short-lived ^{182}Hf–^{182}W chronometer ($t_{1/2} = 8.9$ Ma) is a valuable tool that can effectively constrain the timing of metal–silicate separation and core formation in planetary bodies. Hafnium and W are refractory elements that are strongly fractionated from each other during this process. Hafnium is lithophile and partitions into the silicate phase, whereas W is siderophile and partitions into metallic phases. The application of the ^{182}Hf–^{182}W chronometer to solar system materials, however, is hindered by the effects of neutron capture on W isotopes during exposure to galactic cosmic rays (GCR) (e.g., Leya and Masarik, 2013). Neutron capture reactions cause burnout of ^{182}W, leading to incorrect ^{182}Hf–^{182}W ages. Platinum is an ideal in-situ neutron dose proxy due to the similar neutron capture cross-sections of Pt and W (Leya and Masarik, 2013). Correlating the effects in Pt and W isotopic compositions allows for an estimation of the pre-GCR exposure ε^{182}W value, and consequently, the timing of metal–silicate separation. This method was successfully applied to magmatic iron meteorites by several studies (Kruijer et al., 2014a, 2013a; Wittig et al., 2013).

The IAB meteorites are non-magmatic iron meteorites that have been classified into numerous sub-groups based on their trace-element geochemistry. These include the Main Group (MG), multiple sub-groups including 'Low-Au Low-Ni' (sLL) and 'Low-Au Medium-Ni' (sLM), and several duos or ungrouped samples (Wasson and Kallemeyn, 2002). Recently, nucleosynthetic Mo isotope variations have been used to identify genetic links between these sub-groups (Worsham et al., 2017). That study suggests that while the MG, sLL, sLH and sLM are genetically related, two high-Au sub-groups (sHL and sHH), as well as some ungrouped samples, may derive from a distinct parent body or parent body family.

Trace element data, including highly variable concentrations of siderophile elements, suggest the IABs did not form by simple fractional crystallization in a planetary core, in contrast to magmatic irons (e.g., Wasson and Kallemeyn, 2002; Worsham et al., 2016). Additionally, the IAB irons contain inclusions of chondritic and non-chondritic silicates, sulfides, graphite, and phosphate-bearing

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https://doi.org/10.1016/j.epsl.2017.11.034

Table 1
Sample information and Ir/Pt ratios for analysed IAB meteorites.

Samples	Source	IAB group/ subgroup[a]	Description of inclusions[b]	Ir/Pt[a]	CRE ages (Ma)
Caddo County	ETH	un	ac, nc	0.40	5 ± 1[c]
Canyon Diablo	ETH	MG	gr, ac	0.40	545 ± 40[d]
Cranbourne	ETH	MG	gr	0.34	45 ± 5[e]
Livingstone (TN)	Smithsonian (USNM 1420)	Algarrabo duo	gr	0.17	
Magura	ETH	MG	gr, s	0.37	~250[f]
Odessa	ETH	MG	gr	0.41	875 ± 70[g]
Toluca	ETH	sLL	gr, s	0.43	600 ± 150[e]

[a] IAB classifications and Ir/Pt ratios from Wasson and Kallemeyn (2002). IAB group/sub-groups: MG, main group; sLL, subgroup low-Au low-Ni; un, ungrouped.
[b] Petrographic descriptions of inclusions: ac, angular chondritic; nc, non-chondritic; gr, graphite-rich; s, silicate (Benedix et al., 2000; Buchwald, 1975).
[c] CRE ages from: Takeda et al. (2000).
[d] CRE ages from: Schnabel et al. (2001).
[e] CRE ages from: Chang and Wänke (1969).
[f] CRE ages from: Schulz et al. (2012).
[g] CRE ages from: Voshage and Feldmann (1979).

inclusions (e.g., Benedix et al., 2000; Buchwald, 1975). The angular texture and the olivine- and pyroxene-rich mineralogy of the chondritic silicate inclusions are similar to the winonaite primitive achondrites (Benedix et al., 1998; Bild, 1977). A genetic relationship between the winonaites and MG-IAB iron meteorites is also implied by the identical Mo and O isotope compositions of the two groups, suggesting that they originated from a common parent body (Clayton and Mayeda, 1996; Greenwood et al., 2012; Worsham et al., 2017).

The thermal history and evolution of the IAB parent asteroid, or asteroid family, remains unclear. One theory suggests that the IAB irons formed in melt pools created by impacts into a chondritic parent body or bodies (Choi et al., 1995; Wasson and Kallemeyn, 2002; Worsham et al., 2017, 2016). A competing theory argues that the decay of short-lived radionuclides such as ^{26}Al and ^{60}Fe would produce enough heat for incipient partial melting, which was followed by a catastrophic impact. Subsequent reassembly due to gravity led to the extensive mixing of silicates, sulfides and metals (Benedix et al., 2000; Hunt et al., 2017a; Schulz et al., 2012; Theis et al., 2013). This model can apply to groups with the same genetic affinities (e.g., the MG, sLL, sLH and sLM) as defined by Mo isotopes (Worsham et al., 2017).

Several studies attempted to constrain the history of the IAB parent body using the ^{182}Hf–^{182}W chronometer (e.g., Markowski et al., 2006; Qin et al., 2008; Schulz et al., 2009). The metal–silicate separation ages based on these data and recalculated relative to the most recent value for calcium aluminum rich inclusions (CAI) (ε^{182}W: −3.49 ± 0.07, Burkhardt and Schönbächler, 2015; Kruijer et al., 2014b) range from 3.6 ± 2.1 Ma (Schulz et al., 2009) to 11.5 ± 6.5 Ma after CAI (Markowski et al., 2006). This broad range is partly related to difficulties to correct for neutron capture effects, reasserting the need for a reliable neutron-dose proxy. Recently, attempts to correct for the effects of GCR have yielded ^{182}Hf–^{182}W ages for the IAB irons of between 3.4 ± 0.7 and 6.9 ± 0.4 Ma (Schulz et al., 2012; Worsham et al., 2017). This study also corrects for the effect of GCR on W isotopes in the IABs, using the well-established Pt isotope neutron-dose proxy. We report new W and Pt isotope data obtained on the same sample aliquot for seven samples. The data provide new insights into the thermal evolution of the IAB parent body by constraining the timing of metal–silicate separation. Numerical models of parent body evolution supplement these results. Additionally, the potential of Pt isotopes as an in-situ neutron-dose proxy in iron meteorites is further evaluated, specifically for non-magmatic irons.

2. Samples

Seven IAB meteorites originating from four different IAB groups or sub-groups, as defined by Wasson and Kallemeyn (2002), were chosen for this study. Canyon Diablo, Cranbourne, Magura and Odessa are all assigned to the MG, whereas Toluca is a member of the sLL sub-group, Livingstone (TN) is part of the Algarrabo duo, and Caddo County is ungrouped, but a member of the Udei Station grouplet. These meteorites contain a variety of inclusion types, and cover a range of cosmic ray exposure (CRE) ages (Table 1). All samples are taken from the ETH collection, except Livingstone (TN) (USNM 1420, Smithsonian Institute).

3. Methods

3.1. High-precision Pt and W isotope measurements

Between 1.5 and 3.0 g of each IAB meteorite was prepared for use in this study, and both W and Pt isotope aliquots were taken from the same sample digestion. The sample preparation procedures and Pt and W isotope chemical separations employed in this study are described in the Supplementary Materials. Platinum isotope data for Toluca are taken from Hunt et al. (2017b; Toluca-a).

All Pt and W isotope analyses were performed at ETH Zürich using a Thermo Scientific Neptune *Plus* fitted with a Cetac Aridus II desolvating system and standard H cones. Platinum isotope analyses followed the procedure of Hunt et al. (2017c). Analyses were corrected for instrumental mass bias using the exponential law, and were internally normalized to ^{198}Pt/^{195}Pt (‘8/5’) = 0.2145 (Kruijer et al., 2013a). Samples were analysed with a ^{194}Pt signal of ~2 × 10^{-10} A, equivalent to ~200 ppb Pt, and utilizing ~200 ng Pt per measurement. Each sample was measured relative to the NIST SRM 3140 Pt standard solution, and data are presented in the epsilon notation (i.e., ε^{19i}Pt/^{195}Pt = deviation in parts per 10,000 from the average of the bracketing standards). Repeat analysis of four aliquots of our in-house reference material, the North Chile iron meteorite (IIAB), passed through column chemistry independently give a 2 standard deviation external reproducibility (2 S.D.) of 0.73 for ε^{192}Pt, 0.15 for ε^{194}Pt, and 0.09 for ε^{196}Pt ($n = 19$; Hunt et al., 2017c). Our 2 S.D. external reproducibility for repeat analyses of North Chile is used as the Pt isotope uncertainty throughout this study.

All five W isotopes, along with the interference monitors ^{178}Hf, ^{181}Ta, and ^{188}Os, were measured simultaneously in static mode (Cook and Schönbächler, 2016). Single measurements of each sample were bracketed by measurements of the NIST SRM 3163 W