Siderophile element systematics of IAB complex iron meteorites: New insights into the formation of an enigmatic group

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Abstract

Siderophile trace element abundances and the ^{187}Re - ^{187}Os isotopic systematics of the metal phases of 58 IAB complex iron meteorites were determined in order to investigate formation processes and how meteorites within chemical subgroups may be related. Close adherence of ^{187}Re - ^{187}Os isotopic data of most IAB iron meteorites to a primordial isochron indicates that the siderophile elements of most members of the complex remained closed to elemental disturbance soon after formation. Minor, presumably late-stage open-system behavior, however, is observed in some members of the sLM, sLH, sHL, and sHH subgroups. The new siderophile element abundance data are consistent with the findings of prior studies suggesting that the IAB subgroups cannot be related to one another by any known crystallization process. Equilibrium crystallization, coupled with crystal segregation, solid-liquid mixing, and subsequent fractional crystallization can account for the siderophile element variations among meteorites within the IAB main group (MG). The data for the sLM subgroup are consistent with equilibrium crystallization, combined with crystal segregation and mixing. By contrast, the limited fractionation of siderophile elements within the sLL subgroup is consistent with metal extraction from a chondritic source with little subsequent processing. The limited data for the other subgroups were insufficient to draw robust conclusions about crystallization processes involved in their formation. Collectively, multiple formational processes are represented in the IAB complex, and modeling results suggest that fractional crystallization within the MG may have been a more significant process than has been previously recognized.

1. INTRODUCTION

“Magmatic” iron meteorite groups, such as IIAB, IIIAB, and IVA, are so termed because members of each group have relative and absolute trace element abundances that can largely be accounted for by the fractional crystallization of metallic melts. They have commonly been interpreted to sample the metallic cores of differentiated planetesimals (Lovering, 1957; Fish et al., 1960; Goldstein and Short, 1967; Scott, 1972; Scott and Wasson, 1975). By contrast, the IAB complex is a large group of iron meteorites that is chemically and texturally distinct from the magmatic iron groups.

The members of the IAB complex have major and trace element relationships that cannot be accounted for by simple fractional crystallization (e.g., Scott 1972; Kelly and Larimer, 1977; Wasson et al., 1980; Pernicka and Wasson, 1987; Wasson and Kallemeyn, 2002; Goldstein et al., 2009). Particularly, the ranges in the concentrations of elements that are not strongly fractionated by fractional crystallization, such as Ni, Ge, and Ga, are much greater in the IAB complex than in magmatic groups (**Fig. 1**). Further, some meteorites within the IAB complex contain abundant silicate clasts, which are rare in magmatic groups. Consequently, IAB complex meteorites were once termed “non-igneous”, but are now referred to as “non-magmatic” or “silicate-bearing” (e.g., Wasson, 1970; Willis, 1981; McCoy et al., 1993). The presence of much lower density silicates, most of which are roughly chondritic in both their mineralogy and chemical composition, remains difficult to reconcile with the enveloping, once-molten Fe-Ni metal (Benedix et al., 2000; Wasson and Kallemeyn, 2002). The chemical and isotopic similarity of some silicate inclusions to winonaites, a group of primitive achondrites, has led to the suggestion that at least some IAB meteorites are from the same parent body as the winonaites (Bild, 1977; Clayton and Mayeda, 1996).

The IAB iron meteorite group was initially divided into four separate groups, then consolidated into IAB and IIICD groups, and was later designated group IAB-IIICD as evidence for a compositional continuum accumulated (e.g., Wasson and Schaudy, 1971; Scott et al., 1973; Kracher et al., 1980; Choi et al., 1995). Wasson and Kallemeyn (2002) gathered these meteorites into what is now termed the “IAB complex”, and subdivided it into a main group (MG) and five subgroups based primarily on the clustering of Ni and Au concentrations. The MG comprises the traditional IAB group, whereas the subgroups include the previously defined groups IIIC and IIID. The complex also contains a number of “ungrouped IAB” iron meteorites that bear chemical similarities to the complex, but do not clearly fall into any of the subgroups. It remains unclear what, if any, common processes relate meteorites from the different subgroups, or even whether all IAB complex meteorites come from the same parent body. The processes by which meteorites within a subgroup are related are also debated.

In an attempt to explain the unusual chemical and textural characteristics of the IAB complex, various authors have proposed formation scenarios that invoke melting and/or mixing on either a partially differentiated or undifferentiated parent body. There is, however, currently no consensus on the formation scenario(s) for the IAB complex. Kracher (1985) proposed that the IAB complex irons sample the fractionally crystallized, S-rich core of an internally heated, partially differentiated parent body. By contrast, the trace element modeling of other authors led to the suggestion that the IAB complex originated in impact-generated melt pools in the megaregolith of a chondritic parent body or bodies (Wasson et al., 1980; Choi et al., 1995; Wasson and Kallemeyn, 2002). This scenario can potentially better account for the presence of silicates that are suspended in the metal of some IAB complex irons. Conversely, arguments made by Keil (1997) against the efficiency of impact melting and the longevity of melts led Benedix et al.

(2000) to suggest that radioactive decay was the more likely source of heat than impacts. That study, using evidence from silicate inclusions, proposed a model by which partial differentiation of an internally heated chondritic parent body was followed by impact-induced breakup and reassembly. Studies investigating the petrography and Ar-Ar and I-Xe ages of IAB silicates (Bogard et al., 2005; Vogel and Renne, 2008; Ruzicka and Hutson, 2010), and Hf-W and Pd-Ag model ages of the IAB metals (Schulz et al., 2012; Theis et al., 2013) have supported different aspects of each of the major formational models.

In order to further constrain the formation and crystallization history of IAB complex iron meteorites, we report ^{187}Re - ^{187}Os isotopic systematics, and highly siderophile element abundances (HSE; Re, Os, Ir, Ru, Pt, and Pd are measured here), for bulk samples of 58 IAB complex iron meteorites. The ^{187}Re - ^{187}Os isotope system is used to assess the extent of open-system behavior of the HSE in the IAB complex. The new HSE data are then used to re-examine the classification of meteorites within the complex, and to examine processes that have been previously proposed to relate meteorites within the subgroups. The absolute and relative abundances of the HSE are particularly well suited for this task because their behaviors, resulting from crystal-liquid fractionation and mixing processes, are generally well understood (e.g., Jones and Drake, 1983; Jones and Malvin, 1990; Chabot and Jones, 2003).

2. SAMPLES

All samples examined in this study were obtained from the Smithsonian Institution National Museum of Natural History (**Table 1**). Each of the IAB subgroups is represented by at least three samples. The MG samples examined here represent ~25% of recognized MG iron

meteorites and span the range of MG compositions reported by Wasson and Kallemeyn (2002) (low to high Au contents). Detailed petrographic descriptions of IAB silicates can be found in Bild (1977), Takeda et al. (2000), Benedix et al. (2000), Ruzicka and Hutson (2010), and Ruzicka (2014). Silicate-metal relationships are discussed in Wasson and Kallemeyn (2002). The fraction of IAB iron meteorite samples which contain silicates was estimated by Wasson and Kallemeyn (2002) to be ~ 42% (**Tables 1 and S2** – Buchwald, 1975; McCoy et al., 1993; and Benedix et al., 2000). Silicates occur either as visible inclusions on the mm to cm scale, or as finely disseminated grains on the order of 100 μm across (Wasson and Kallemeyn, 2002). Abundances of silicate clasts in silicate-bearing IAB irons can range from a few vol. % to 40-60 vol. % (e.g., McCoy et al., 1996; Tomkins et al., 2013). For example, in Hope, Bogou, and Smithville, silicates were observed as rare phases in association with troilite and graphite (and only as a single grain in Hope) (Buchwald, 1975). In contrast, Landes and Campo del Cielo, have abundant silicate clasts (e.g., Bunch et al., 1972; Tomkins et al., 2013). The majority of the silicates are broadly chondritic in terms of both their chemical composition and mineralogy, though Benedix et al. (2000) also described sulfide-rich, graphite-rich, and phosphate-rich, non-chondritic silicates. In terms of the metal, many IAB complex irons display a Widmanstätten pattern upon etching with dilute acid, with kamacite and taenite lamellae ranging from sub-mm to mm scale. Samples from the IAB complex represent the range of textural types present in iron meteorites from hexahedrite to ataxite (Benedix et al., 2000).

3. ANALYTICAL METHODS

3.1. HSE chemistry

Highly siderophile element concentrations and Os isotopic compositions were obtained using methods previously described (Cook et al., 2004; Walker et al., 2008; McCoy et al., 2011). The meteorites were cut into 30 to 300 mg pieces using a water-cooled *Leco “Vari-cut”* saw with a 12.7 cm diamond wafering blade, which was cleaned with carborundum between each sample. For some silicate-bearing meteorites (e.g., Landes and Woodbine), it was impossible to separate pure metal from silicate for digestion, although silicate inclusions were avoided as much as possible. All samples were mechanically abraded using carborundum paper, and sonicated in ethanol three times to remove any adhering rust, fusion crust, or saw blade contamination. Masses of the samples digested are given in **Table 1**. Each sample was spiked prior to digestion with a combined platinum-group element (PGE) spike (^{191}Ir , ^{99}Ru , ^{194}Pt , and ^{105}Pd) and a combined Re-Os spike (^{185}Re and ^{190}Os) for isotope dilution. Occasionally, it was necessary to spike with an additional individual ^{105}Pd spike to supplement the combined PGE spike because it was too dilute in ^{105}Pd for some of the Pd-enriched samples.

Samples and spikes were digested and equilibrated in *Pyrex*[®] Carius tubes at 220°C for at least 24 hours, using 5 ml of concentrated HNO_3 and 2.5 ml of concentrated HCl . The metal phase was completely digested using this technique, but any silicates present were not dissolved. Upon opening the tubes, oxidized volatile Os was immediately extracted into carbon tetrachloride (CCl_4) and then back-extracted and reduced using concentrated HBr (Cohen and Waters, 1996). The Os was further purified via microdistillation (Birck et al., 1997). The Os total analytical blank was negligible for all samples, comprising $< 0.1\%$ of the total Os extracted, averaging 4 ± 2 pg, with an isotopic composition of $^{187}\text{Os}/^{188}\text{Os} = 0.15 \pm 0.04$ ($n=7$).

The HSE remaining in the aqueous phase following Os extraction were separated and purified using anion exchange chromatography (Rehkämper and Halliday, 1997). Samples were

centrifuged to separate and allow removal of silicate residue. Each sample was loaded onto a disposable 2 ml Biorad column filled with 1.7 ml of AG1X8 100-200 mesh resin in 1M HCl. After most non-analytes were rinsed off the column with 1M HCl, 0.8M HNO₃, and 1M HCl-1M HF, Re and Ru were eluted in 6M HNO₃, followed by Pt and Ir in concentrated HNO₃. Finally, Pd was eluted in concentrated HCl. Rhenium and Ru cuts were further purified using a smaller version of the primary anion column that was made using disposable transfer pipettes and 250 µl of resin. The Re-Ru, Pt-Ir, and Pd cuts were then evaporated to dryness and re-dissolved in a small amount of *aqua regia* to reduce resin-derived organics. The HSE cuts were dried again and taken up into solution in 0.8 M HNO₃ for analysis. The total analytical blanks for Re, Ir, Ru, Pt, and Pd were 1 ± 2 , 3 ± 10 , 30 ± 40 , 300 ± 600 , and 40 ± 80 pg, respectively (n=7). These blanks were typically less than 1% of the total element analyzed, with the exception of the Pt blank, which accounted for 2-3% of the Pt separated from some lower abundance samples (i.e., Mount Magnet, Freda, Dayton, Edmonton (KY)). The Re blank accounted for $\leq 0.1\%$ of the total Re extracted for nearly all samples. The exceptions are Dayton and Freda, where the blank accounted for 0.2% and 0.9% of the total Re, respectively. All blanks are sufficiently low as to have an inconsequential effect on the isotope systematics.

3.2. Mass spectrometry

Osmium was analyzed using a *VG Sector 54* thermal ionization mass spectrometer in negative ion mode (N-TIMS). Loading procedures were similar to those reported by Cook et al. (2004). Samples were dissolved in concentrated HBr and 1 to 50 ng of Os were loaded onto an outgassed Pt filament. The sample was dried by applying a 0.5 A current through the filament. Once dry, ~ 0.8 µl of a saturated solution of Ba(OH)₂ was deposited on top of the sample and

used as an electron activator. For samples and standards, the Os isotope beams were collected in multiple Faraday cups using a static analytical routine. The external reproducibility of $^{187}\text{Os}/^{188}\text{Os}$ for an Os reference material was $\pm 0.03\%$ (2σ). For samples, however, the average internal uncertainty on the $^{187}\text{Os}/^{188}\text{Os}$ measurement was $\pm 0.04\%$ ($2\sigma_m$). Because the blank correction for $^{187}\text{Os}/^{188}\text{Os}$ was negligible, we assigned an uncertainty of $\pm 0.04\%$ to $^{187}\text{Os}/^{188}\text{Os}$. In cases where lower signal intensity made the statistics of the individual measurement $> 0.04\%$, the standard error of the mean of the measurement is reported ($2\sigma_m$) (low-abundance samples Freda and Mount Magnet). The uncertainties for Os concentrations are estimated to be $\pm 0.1\%$.

Rhenium, Ir, Ru, Pt, and Pd concentrations were determined using a *Nu Plasma* multi-collector inductively coupled plasma mass spectrometer (MC-ICP-MS). Sample solutions were introduced into the plasma using a *Cetac Aridus*TM desolvating nebulizer. Analyses were conducted in static mode using multiple Faraday cups. Spiked ratios were measured to better than $\pm 0.1\%$ (2σ). For Ir, Ru, Pt, and Pd, instrumental mass bias was corrected using a fractionation factor obtained by comparing isotope ratio measurements of standards with the accepted isotope ratios. Rhenium cuts were doped with W in order to correct for instrumental mass bias, normalizing to $^{184}\text{W}/^{186}\text{W} = 1.0780$ (Kleine et al., 2004). The uncertainties of Ir, Ru, Pt, and Pd concentrations were estimated to be better than $\pm 3\%$, based on repeated analyses of spiked aliquots of previously well-characterized iron meteorites which were used as reference materials (Walker et al., 2008). Uncertainties for Re concentrations are estimated to be $\pm 0.18\%$. For $^{187}\text{Re}/^{188}\text{Os}$ ratios, the uncertainties were estimated to be $\pm 0.2\%$.

4. RESULTS

4.1. ^{187}Re - ^{187}Os isotopic systematics

Rhenium-187 decays to ^{187}Os via β^- decay with λ for $^{187}\text{Re} = 1.666 \times 10^{-11} \text{ yr}^{-1}$ (Smoliar et al., 1996). The siderophile element systematics of meteorites that do not plot on a primordial ^{187}Re - ^{187}Os isochron must be considered with caution, as non-isochronous samples may reflect post-formation gain or loss of Re and/or Os, and likely other HSE. Rhenium-Os isotopic data are presented in **Table 1** and **Fig. 2**. The moderate fractionation of Re from Os in the MG produces a range in $^{187}\text{Re}/^{188}\text{Os}$ from 0.346 (Magura) to 0.909 (Hope). The corresponding range in $^{187}\text{Os}/^{188}\text{Os}$ is 0.12270 to 0.16727. If the other subgroups are included in the range, the lowest $^{187}\text{Re}/^{188}\text{Os}$ and $^{187}\text{Os}/^{188}\text{Os}$ ratios are for Carlton (sLM) and the highest are for Georgetown (ungrouped).

Rhenium-Os isotopic data were regressed for the MG and individual IAB subgroups using ISOPLLOT (Ludwig, 2003) (**Table 2**). Because it is represented by the largest number of samples in this study and in the IAB complex, the MG isochron is used for comparison to the other subgroups. The MG isochron gives an age of $4574 \pm 20 \text{ Ma}$ and an initial $^{187}\text{Os}/^{188}\text{Os}$ of 0.09532 ± 0.00017 (MSWD = 4.2). This age and initial $^{187}\text{Os}/^{188}\text{Os}$ overlap, within uncertainties, of the composite group IAB age and initial ratio of 4537 ± 21 and 0.09556 ± 0.00016 reported by Horan et al. (1998). The age and initial ratio are also similar to results for magmatic iron meteorite groups (Smoliar et al., 1996; Shen et al., 1996; Walker et al., 2008; McCoy et al., 2011). The isochrons for most of the individual subgroups yield ages and initial $^{187}\text{Os}/^{188}\text{Os}$ that overlap with those of the MG, although the precision of the regressions vary considerably, in part due to the limited range in Re/Os for some subgroups (sLH, sHH). The sHL isochron yields an apparent age ~100 Myr younger than isochrons for the MG, sLL, and sLM subgroups; however,

it was generated with data from only three meteorites, and this age is not considered to be meaningful.

Part in 10,000 deviations of data for individual samples, relative to the slope and initial $^{187}\text{Os}/^{188}\text{Os}$ of the MG isochron, are given as Δ_{Os} values in **Table 1** (Walker et al., 2002). The Δ_{Os} values are calculated by $\Delta_{\text{Os}} = 10^4(^{187}\text{Os}/^{188}\text{Os}_{\text{IAB}} - (0.09532 + 0.07918 \times ^{187}\text{Re}/^{188}\text{Os}_{\text{IAB}}))$, where 0.09532 is the initial $^{187}\text{Os}/^{188}\text{Os}$ and 0.07918 is the slope of the MG isochron (**Table 2**). The 13 samples of the 58 IAB meteorites examined here that do not plot within uncertainty of the MG isochron are Carlton and Lamesa (sLM); Tazewell, Dayton, and Freda (sLH); Chebankol (sHL); ALHA 80104 and Mount Magnet (sHH); and Pitts, EET 84300, Mundrabilla, Waterville, and Georgetown (ungrouped).

$^{187}\text{Re}/^{188}\text{Os}$ and $^{187}\text{Os}/^{188}\text{Os}$ ratios for duplicate analyses of an individual sample are typically beyond uncertainties of each other. For instance, Shrewsbury_A (subsample piece A) has $^{187}\text{Re}/^{188}\text{Os}$ and $^{187}\text{Os}/^{188}\text{Os}$ of 0.3038 ± 0.006 and 0.11949 ± 0.0005 , whereas those of Shrewsbury_B are 0.2995 ± 0.006 and 0.11912 ± 0.0005 . The only samples for which both $^{187}\text{Re}/^{188}\text{Os}$ and $^{187}\text{Os}/^{188}\text{Os}$ are within uncertainty of each other are Sardis and Edmonton (KY). Despite the variations in the ratios, most duplicate analyses plot within analytical uncertainties of the MG isochron.

4.2. HSE data for bulk samples

Highly siderophile element concentration data for bulk samples are also presented in **Table 1**. A detailed comparison of the data presented here with literature data is provided in the supplementary materials (**Table S3**). Generally, our new HSE data are within $\pm 20\%$ of values

reported by Goldberg et al. (1951), Smales et al. (1967), Crocket (1972), Pernicka and Wasson (1987), Hirata and Masuda (1992), Hoashi et al. (1993), Choi et al. (1995), Shen et al. (1996), Horan et al. (1998), and Wasson and Kallemeyn (2002). Some HSE data for certain samples differ by > 30% from literature values. These are the MG samples Sarepta (Re and Ir) and Cranbourne (Re, Os, Ir, Ru, Pt, and Pd); sLL sample Toluca (Ru); sLM samples Persimmon Creek (Re), Anoka (Pt), Mungindi (Re and Pd), Carlton (Pd), and Lamesa (Pd); sLH samples Freda (Re and Ir) and Dayton (Pd); sHH sample Mount Magnet (Os and Ir); and ungrouped sample Woodbine (Pt). Rhenium data reported here are systematically lower than those reported by Wasson and Kallemeyn (2002), but no other HSE have systematic offsets from the literature.

Fourteen of the IAB complex iron meteorites were processed multiple times to assess sample heterogeneity and/or analytical reproducibility. Duplicate analyses, denoted in **Table 1**, yielded variable concentrations, with most duplicates having percentage differences of < 3 %, but ranging up to ~10-20%. The HSE ratios of duplicates also varied, ranging from < 3% to 23% (**Table S4**). Typically, duplicate analyses that were characterized by large variability in absolute HSE abundances were also characterized by HSE ratios that were variable.

We attribute the elemental and isotopic heterogeneities in bulk samples to primarily reflect chemical heterogeneity at the sample size processed. The heterogeneities observed most likely reflect variable distributions of kamacite and taenite in different pieces, as most of the HSE are fractionated between the two metal phases (e.g., Hirata and Nesbitt, 1997). The differences highlight the difficulty by which comparatively small pieces of a IAB meteorite can be precisely characterized by a single “bulk” composition. Nevertheless, the scale of the sample heterogeneity is not sufficient to influence the conclusions drawn below that are based on large differences in absolute and relative HSE abundances within and between subgroups.

Overall, HSE abundances in samples from the IAB complex range over as much as three orders of magnitude (**Fig. 1**), although concentrations within subgroups span no more than two. Palladium abundances vary less than the other HSE, both between and within subgroups, having a range of < 50% in each of the subgroups. The chondrite-normalized HSE patterns, plotted from left to right in order of decreasing 50% condensation temperatures, are characterized by similar features within a given subgroup that are not shared between all subgroups (**Fig. 3**). Unlike in magmatic iron groups, HSE patterns for some meteorites cross in most of the subgroups. Also, most subgroups have HSE concentrations that are fractionated relative to chondritic abundances. The exception is the sLL subgroup, which has suprachondritic absolute HSE abundances which are in approximately chondritic relative abundances.

The HSE patterns of some subgroups resemble the patterns of some magmatic iron meteorite groups. Of note, the MG patterns share some similarities with the IVA group, whereas the sHL and sHH subgroups have HSE patterns that are broadly similar to late-crystallized IIAB and IIIAB iron meteorites (**Fig. 3** – Pernicka and Wasson, 1987; Hoashi et al., 1993; Cook et al., 2004; McCoy et al. 2011). By contrast, the sLM and sLH subgroups have distinctive HSE patterns, compared to known magmatic irons, and are characterized by absolute Re, Os, Ir, Ru and Pt abundances that range from broadly chondritic to significantly subchondritic, but in roughly chondritic relative proportions. Palladium is considerably enriched, relative to chondrites, and the other HSE in these subgroups.

The ungrouped samples are characterized by HSE patterns that are highly variable. This is expected, as these samples do not belong to any one subgroup (**Fig. 4**). Some ungrouped members of the IAB complex share similarities in HSE patterns with certain IAB subgroups. Of

note, the HSE pattern of Caddo County is similar to the sLL subgroup, and the pattern of Pitts is similar to the sLM subgroup.

5. DISCUSSION

5.1. Re-Os isotopic system

Because most of the IAB samples plot on or near a primordial isochron (**Fig. 2**), the ^{187}Re - ^{187}Os isotope systematics require that the IAB complex was, in large part, a closed system to the movement of siderophile elements after the first several 10's of Ma after Solar System formation, and therefore, that their HSE abundances reflect those at the time of crystallization. Conversely, the samples that do not plot within uncertainties of the MG isochron have absolute and relative HSE abundances that may reflect later-stage processes, either on the parent body, or as a result of terrestrial alteration. Late-stage parent body processes, which likely occurred greater than ~ 40 Myr after the crystallization of the non-isochronous samples, could be related to thermal or shock metamorphism due to impacts, though shock effects in IAB iron meteorites are minor (Benedix et al., 2000). Alternatively, if a late impactor was volatile-rich, aqueous alteration could also modify the Re/Os. A few samples studied here showed evidence of terrestrial weathering as extensive rust, including Ballinger and Morasko (MG); Deport and Shrewsbury (sLL); Persimmon Creek (sLM); TIL 91275, Mertzon, and Woodbine (ungrouped). Of these, none are resolvable from the MG isochron, suggesting that the presence of rust is not indicative of non-isochronous behavior.

The non-isochronous samples are largely confined to the sLM, sLH, sHL, and sHH subgroups. This suggests that alteration on the parent body or Earth was either more pronounced

in these samples, because of their generally lower Re and Os abundances, relative to the MG, or that the meteorites from these subgroups were subject to more intense alteration than those in other subgroups. Implicit in the second scenario is the assumption that the alteration was localized, or that the sLM, sLH, sHL, and sHH subgroups were spatially separated from the other subgroups, perhaps in different portions of a single parent body, or on distinct parent bodies. Nevertheless, HSE mobility in the complex was evidently limited because none of the samples plot more than 2% off of the MG isochron. Therefore, the conclusions drawn below are likely unaffected by the minor open-system behavior.

5.2. Classification within the IAB complex

The IAB complex is classified as such due to an observed continuum in taxonomically useful elements. In general, magmatic and silicate-bearing iron meteorite groups have been classified based on Ga, Ge, Ni, Au, and Ir abundances (e.g., Goldberg et al., 1951; Lovering et al., 1957; Wasson, 1967; Wasson, 1969). The concentrations of Ga, Ge, and Ni are not strongly affected by crystallization processes and have proven to be important in distinguishing among magmatic iron meteorite groups that underwent extensive crystal-liquid fractionation processes (Goldberg et al., 1951; Lovering et al., 1957). Conversely, the IAB complex is characterized by a much larger range in Ni than any magmatic iron group, accompanied by nearly as great a range in Ir and Os concentrations as in the most highly fractionated magmatic groups (**Fig. 1**). Further, the IAB complex is characterized by siderophile element concentrations that define broad clouds of data on x-y plots, in comparison to the more sharply defined trends characteristic of the magmatic groups. In contrast to the complex as a whole, the IAB subgroups have considerably

more restricted ranges in, e.g., Ni and Os. The new HSE data provide no cause for change in the conclusion of Wasson and Kallemeyn (2002) that the collective members of the IAB complex cannot be related by a simple crystal-liquid fractionation process. Thus, if the subgroups are all from the same parent body, it must be concluded that the subgroups crystallized from distinct melts, presumably in different locations.

Wasson and Kallemeyn (2002) based their division of the IAB complex into subgroups on the clustering of data when plotting Au versus other siderophile elements. The extended database for HSE here offers the opportunity to re-examine the subgroups using elements with somewhat different volatilities and crystal-liquid fractionation behaviors. The new high precision Pd data are particularly useful for this because, compared to Au, Pd has a nearly $\sim 300^\circ$ higher 50% condensation temperature (at 10^{-4} bar), yet is only slightly less incompatible in solid metal-liquid metal systems. Thus, Pd can be strongly modified relative to Au by processes that can lead to the formation of metals, but should behave similarly to Au, once a liquid-solid metal system is established.

For the MG and each of the subgroups, Pd is well correlated with both the Ni and Au concentrations reported by Wasson and Kallemeyn (2002) (**Fig. 5**), strengthening the case that the chemical basis of the subgroup divisions is robust. In order to differentiate the subgroups that show overlap in the classification of Wasson and Kallemeyn (2002) (MG and sLL), or that are characterized by similar HSE patterns (sLM and sLH; sHL and sHH), each of the subgroups are shown on a plot of Au vs. Pd (**Fig. 5**). In Au and Pd compositional space, the sLL subgroup is characterized by some overlap with the MG, but this overlap is reduced on a plot of Ni vs. Pd.

The new HSE data can be used to reassess some of the classifications of Wasson and Kallemeyn (2002) for particular meteorites (**Figs. 3-5**). For example, Osseo and Sardis, which are termed “unclassified”, are included in the IAB complex, but are not classified as ungrouped or belonging to a specific subgroup; however, both of these samples plot within the MG fields on our new HSE variation diagrams. In Wasson (2011), Osseo is referred to as a MG iron. Therefore, we will assume both meteorites are MG irons.

Morasko and the chemically similar irons Seeläsgen, Burgavli, and Sarepta have Pd abundances within the range of the MG, and these irons were classified into the MG by Wasson and Kallemeyn (2002). Conversely, Pilski et al. (2013) recommended that Morasko and chemically similar irons be considered distinct, anomalous members of the MG, based on their low Ir concentrations. However, Ir is strongly incompatible in metallic systems with appreciable S, and splitting vs. lumping in this case could lead to an incomplete picture of the crystallization of the MG. Therefore, we continue to assume that irons with similar Pd, but low Ir are members of the MG, as discussed below. Additionally, Morasko and Seeläsgen are likely paired (Kracher et al., 1980; Pilski et al. 2013), and the similarity of the new HSE data for both meteorites further supports a Morasko-Seeläsgen link.

Three ungrouped IAB irons, Caddo County, Zagora, and TIL 91725 (**Fig. 4**), which fall into the “Udei Station grouplet” defined by Wasson and Kallemeyn (2002), have similar HSE patterns and abundances to the sLL subgroup and plot within or very near sLL fields on Pd vs. other HSE plots. Wasson and Kallemeyn (2002) suggested that the Udei Station grouplet is closely related to the sLL subgroup and the new HSE data are consistent with this although Ni abundances reported by Wasson and Kallemeyn (2002) are ~15 % higher in Caddo County and Zagora than the average Ni concentration of the sLL subgroup.

Pitts and Woodbine (Pitts grouplet) have higher total HSE abundances than the sLM subgroup and do not lie on some of the linear arrays defined by the sLM subgroup on HSE plots (**Fig. 5**). However, the sLM subgroup and Pitts and Woodbine share the characteristic of enriched Pd relative to the other HSE, which is suggestive of a similar, but likely not common, heritage. In the sLH subgroup, the considerably higher Pd abundance in Freda suggests that it is not related to Tazewell and Dayton.

The small number of samples available for analyses in the sHL and sHH subgroups, coupled with the scatter in the sHL and sHH fields (**Fig. 5**), make it difficult to assess relations within these groups (Wasson, 2011). However, Sombroete, tentatively assigned to the sHL subgroup by Wasson and Kallemeyn (2002), contains silicate with distinct $\Delta^{17}\text{O}$, compared with the small range of $\Delta^{17}\text{O}$ values in other IAB complex meteorites (Clayton and Mayeda, 1996). Further, Sombroete is unusually P-rich, and its silicate inclusions are fractionated with respect to their major and trace lithophile element abundances, relative to the commonly chondritic silicates in other IAB irons (Ruzicka et al. 2006). Ruzicka et al. (2006) considered Sombroete to be an ungrouped iron, not related to the IAB complex, and the HSE data are consistent with this interpretation. Quarat al Hanish and Chebankol have Pd, Ni, Au, and Ga abundances that bracket those of Sombroete, so if Sombroete is unrelated to the other sHL iron meteorites, Quarat al Hanish and Chebankol have little in common with one another. In the sHH subgroup, two of the three irons we analyzed have nearly identical HSE patterns, whereas the third, Mount Magnet, has a much more fractionated pattern. It is, therefore, possible that several sHL and sHH irons represent different parental melts.

5.3. Crystal-liquid fractionation within the IAB MG

Despite the great chemical diversity within the complex, individual IAB subgroups are characterized by more modest, yet in some subgroups significant ranges in siderophile element concentrations. A major question about the IAB complex is then whether or not meteorites within a subgroup can be related to one another by crystal-liquid fractionation processes or some other process, such as mixing. Using the new HSE data for meteorites from each of the subgroups, we re-examine the potential for equilibrium and fractional crystallization, as well as crystal-liquid mixing to have produced the chemical variability observed in the individual subgroups.

Solid metal-liquid metal fractionation, such as fractional crystallization, can produce large chemical variations among the HSE, as observed in magmatic iron meteorites (**Fig. 1**). The generation of robust crystal-liquid fractionation models requires knowledge of the partitioning behaviors of the HSE in metallic systems. These behaviors are highly dependent on the initial composition of the melt, particularly with respect to the concentrations of the light elements S, P, and C. The light elements normally behave incompatibly, and so become increasingly enriched in the liquid phase as metallic iron systems evolve. As the concentrations of these elements increase in metallic systems, the general tendency for most of the HSE is to partition more readily into the solid phase (e.g., Willis and Goldstein, 1982; Jones and Drake, 1983; Jones and Malvin, 1990). When the light element contents are taken into account, crystallization of a metallic melt will fractionate the HSE between solid and liquid metal in predictable and systematic ways. Rhenium, Os, Ir, and Ru normally behave compatibly, Pt can behave either compatibly or incompatibly, and Pd and Au typically behave incompatibly (e.g., Chabot and Jones, 2003).

For the modeling conducted here, the solid metal-liquid metal D values were calculated using a “parameterization method” which uses empirical parameters derived from experimental partitioning data (supplementary material; Jones and Malvin, 1990; Chabot and Jones, 2003; Chabot et al., 2014). In brief, the parameterization depends on the mole fractions of S, P, and C (X_S , X_P , and X_C); an “Fe domains” variable, which constitutes the availability of light element-free domains into which siderophile elements prefer to partition; and β , which is the slope of the fit to experimental data for a given element on a plot of the natural log of D vs. the natural log of Fe domains. Because the IAB complex was characterized by comparatively high C, the parameterization method and modelling were modified to account for C (see SM section 4). The main input variables used to calculate D values are, therefore, the initial abundances of S, P, and C. The initial abundances of these light elements were chosen based on a combination of the literature estimates of their concentrations derived from modal mineralogy, and the concentrations which generated D values that can best account for the range of HSE concentrations in the individual subgroups.

As the largest and most chemically complex IAB group, the MG is considered first. Previous estimates of initial S contents in IAB melts are comparatively high, and are based mainly on modal observations pertaining to MG irons, with concentrations ranging from < 9.7 wt. % S, to just below the eutectic composition of the FeNi-FeS system (~31 wt. % S) (Kracher, 1985; McCoy et al., 1993; Wasson and Kallemeyn, 2002). Wasson and Kallemeyn (2002) also estimated P contents to be roughly 0.2 times those of S, based on modal abundance estimates of P and S from high-S MG meteorites. Carbon was estimated to be ~ 0.2 wt. % (e.g., Moore et al., 1969; Buchwald, 1975; Wasson and Kallemeyn, 2002).

5.3.1 Crystal segregation modelling of the MG

Wasson and Kallemeyn (2002) proposed a model for the formation of the MG IAB irons based on crystal segregation (CS) to account for most of the chemical variations observed in the suite. Their model begins with impact induced melting of a megaregolith on a porous chondritic parent body to produce a metallic liquid with broadly chondritic relative abundances of siderophile elements. This step is followed by partial equilibrium crystallization of that liquid. The model then assumes the metal crystals are separated from the melt, and the crystals and liquid subsequently re-mixed in varying proportions. The proposed two-component mixing scenario can account for the linear trends defined by siderophile element plots for the MG, such as Au vs. Ni and Au vs. Ga. In the CS model, the liquid metal is assumed to crystallize sufficiently quickly to prevent the separation of metal and silicates that were presumably derived largely from the target, accounting for the presence of silicates in some MG irons. It is also assumed that chemical equilibrium is retained between the crystals and metallic liquid as crystallization proceeds.

To re-examine the Wasson and Kallemeyn (2002) model with the new HSE data, we apply the equilibrium crystallization equation (1) to calculate the concentration of a given element in the liquid (C_L) and solid (C_S ; where $C_S = D \cdot C_L$) using the initial bulk concentration of the system (C_0), the fraction of crystallized solid (F), and D values where:

$$C_0 = F \cdot D \cdot C_L + (1 - F)C_L \quad (1)$$

Wasson and Kallemeyn (2002) calculated the solid and liquid endmember compositions by assuming 35% crystallization ($F = 0.35$) before the separation of metal crystals and metallic liquid, and choosing D values which would correspond to the range of abundances observed in

the MG. These authors assumed a D_{Ir} of 3.0. To be consistent, we also assumed 35% crystallization. For our CS model, initial S, P, and C contents were iteratively chosen by applying the parameterization of Chabot and Jones (2003) and Chabot et al. (2014) so that D_{Ir} was also 3.0 (IP- MG_{CS}; **Table 3**). The parameterization was then used to calculate the corresponding D values for the other HSE (**Table 4**). As in Wasson and Kallemeyn (2002), the concentrations of the HSE in the initial system were chosen to be the average of the concentrations present in the MG, excluding low Ir members of the MG. By combining the estimated initial concentrations with the appropriate values for D, and assuming 35% crystallization, equation 1 was used to calculate HSE concentrations in the liquid and solid endmembers (**Table 5**). Two-component mixing calculations were then used to generate the mixing line with compositions incorporating 0% to 80% of the solid fraction. The silicates do not have any compositional effect on the model because of their considerably lower HSE concentrations relative to the metal phases.

The CS model works well to describe the linear trends defined by Re vs. Os, Ir, and Pd (**Fig. 6**). However, because plots of Re vs. Ru and Pt do not produce linear trends ($r^2 < 0.1$), the CS model alone cannot reproduce the data (**Fig. 6**). The scatter in Ru and Pt persists when Au data from Wasson and Kallemeyn (2002) are plotted vs. Ru and Pt, so the scatter is not a result of plotting these elements versus the compatible element Re, rather than the incompatible element Au.

An additional problem with the CS model is that, at least using similar D values as Wasson and Kallemeyn (2002), it cannot account for MG meteorites with low Re, Os, and Ir concentrations (e.g., Hope, Morasko, Sarepta, and Burgavli). Wasson and Kallemeyn (2002) and Pilski et al. (2013) suggested that the iron meteorites with low Re and Ir may have crystallized as

a result of local, volumetrically minor fractional crystallization. Choi et al. (1995) also suggested that minor fractional crystallization and mixing contributed to the scatter observed in some trace element distributions in the IAB complex. Consequently, in the next section we assess the effects of late-stage fractional crystallization on this system.

5.3.2 Crystal segregation-fractional crystallization hybridized model

The presence of silicate inclusions in many of the MG members has been cited to argue against fractional crystallization (Wasson et al., 1980; Wasson and Kallemeyn, 2002). This argument has been made because fractional crystallization may operate over long periods of time, during which newly formed Fe-Ni crystals are likely physically and chemically removed from the liquid. At the same time, the liquid will be continuously homogenized. Lower density silicates are highly buoyant and, therefore, likely cannot remain entrained in metal on the longer time scales necessary for homogenizing and mixing the melt (e.g., Tomkins et al. 2013).

Although silicates are common in the MG irons with high total HSE abundances, silicates are rare or absent in most of the more chemically fractionated members of the MG (silicate occurrences are noted in **Tables 1** and in more detail in **Table S2**). For this reason, and because the HSE patterns for the more fractionated members of the MG are broadly similar to patterns produced by extensive fractional crystallization in some magmatic iron groups (**Fig. 3a**), an additional stage of fractional crystallization is assessed as a means of generating the chemical variations in the MG that cannot be accounted for by the CS model alone (e.g., the low Re-Os-Ir abundance samples). This is termed the “CS-FC model”.

Several variables in the CS model were modified for our CS-FC model. The degree of crystallization before the solids and liquids were separated was set at 0.15, rather than the 0.35 used for the CS model. This was done so that the CS portion of the new model, which used higher D values, would lead to the production of a range of HSE concentrations that included Osseo, but was similar to the lower range achieved by the Wasson and Kallemeyn (2002) model. For the same reason, our CS-FC model only involves mixtures incorporating 0% to 60% of the solid endmember, as compared to 0-80% solid in the CS model. Another important difference between the CS model and the CS-FC model is that the initial bulk composition of the CS-FC model was taken as the mean of the MG, with the exception of both the low Ir samples and the high Pt samples (**Table 5**). This was done because, on plots of Re vs. Pt and Ru, the low Pt and Ru samples define a generally linear correlation with Re. Therefore, by using the lower-Pt mean, the CS model can be fit to the samples that form a linear feature within the scatter.

To model fractional crystallization subsequent to CS, the liquid endmember composition generated by the modified CS model was used as the initial liquid composition, from which the evolving solid and liquid compositions produced by fractional crystallization were calculated. A number of studies have successfully modeled the effects of fractional crystallization on siderophile element abundances to account for the compositional ranges observed in the major magmatic iron meteorite groups (Wasson, 1999; Wasson and Richardson, 2001; Wasson et al., 2007; Walker et al., 2008; McCoy et al., 2011). The fractional crystallization modeling conducted here generally follows the approach that Walker et al. (2008) and McCoy et al. (2011) used to model the IVB and IVA magmatic iron meteorite systems, respectively. Although the interpretation that the IVA group was fractionally crystallized has been called into question by Albarède et al. (2013), the fractional crystallization model of McCoy et al. (2011) can well

account for the many of the siderophile element variations in the IVA group. As opposed to the equilibrium crystallization in the CS model, the fractional crystallization portion of the model must account for the systematic way with which D values for the HSE change with the concentration of the incompatible S, P, and C in the crystallizing liquid.

A modified fractional crystallization model is applied, using a constant D_S of 0.005, to iteratively calculate X_S , X_P , and X_C , as well as the corresponding D values and β_{SPC} for each of the HSE. The β_{SPC} represents the combined β values for S, P, and C, and collectively accounts for the effects of these elements. The concentrations of the HSE in the evolving liquid are calculated for each 0.1% step of crystallization (n) using the equilibrium crystallization equation (1) above, but by iteratively changing F; the light element abundances and, thus, the D values; and by equating C_0 with the C_L at each prior step (n-1), we account for the fractional nature of the crystallization, such that:

$$C_{Ln} = C_{Ln-1} / (F_{n-1} + 1 - F_{n-1} D_n) \quad (2)$$

The solid composition is calculated by $C_{Sn} = C_{Ln} D_n$ at each step.

Several combinations of S, P, and C abundances were trialed in an attempt to find the set of D values that generates models which best fit the observed HSE trends. In light of the constraints from the literature outlined above, our preferred model assumes initial S, P, and C contents of 10, 2, and 0.05 wt. %, respectively (IP-MG_{CS-FC}; **Table 3**). The corresponding initial D values for the HSE are reported in **Table 4**.

In general, iron meteorites that form by fractional crystallization can sample a solid that was in equilibrium with its associated liquid, a liquid composition trapped within earlier formed solids, or a mixture of the two (Wasson, 1999). Fractionally crystallized iron meteorites,

therefore, tend to plot along a solid track that traces the chemical evolution of precipitating solids, a liquid track that traces the chemical evolution of the liquid, or along mixing lines between the two (Cook et al., 2004; Wasson et al., 2007; Walker et al., 2008).

The results of the preferred CS-FC model are shown in plots of Re or Pt vs. other HSE, or HSE ratios (**Fig. 7**), and in plots of Au vs. other HSE and moderately siderophile elements reported by Wasson and Kallemeyn (2002), in the supplementary materials. Rhenium/Os and Pt/Os ratios are useful to plot because it is assumed that the metals examined were ultimately derived from a body with a bulk chondritic composition. Thus, it is presumed that the initial metallic melts also had Re/Os and Pt/Os within the range of bulk chondrites. This range is well established through present day $^{187}\text{Os}/^{188}\text{Os}$ and $^{186}\text{Os}/^{188}\text{Os}$ of bulk chondrites (Walker et al., 2002). Hope is the most fractionated sample in the MG and is denoted in **Figure 7**, along with Osseo, the least fractionated. We also label the data for Morasko and Magura, two moderately fractionated samples, in each plot. Because the CS model is a mixing model, samples that do not plot between the solid and liquid tracks but near the CS mixing line (thick black line in **Fig. 7**), are considered to be accounted for by the CS portion of the CS-FC model. In the Re vs. Re/Os plot (**Fig. 7a**), all MG samples plot between the solid and liquid tracks, or near the CS mixing line. On Pt vs. Pt/Os and Re vs. Ru, Pt, and Pd plots, most MG samples plot either on or near the solid track, between the solid and liquid tracks, or near the CS mixing line. In Re vs. Re/Os and Pt vs. Pt/Os compositional space (**Fig. 7a-b**), the Re/Os and Pt/Os of the liquid endmember from the CS model are slightly higher than chondritic. However, because the initial melt in the CS model is the bulk composition of the MG, the condition of chondritic Re/Os can be met. The Pt/Os of the MG average remains slightly suprachondritic (**Table 5**). Therefore, the CS-FC

model provides good fits for the observed HSE variations in the MG which cannot be accounted for with the CS model alone.

The observation that the hybrid CS-FC model is in good agreement with most of the observed data for the MG is not without caveats. First, we note that the initial S, P, and C contents provide non-unique solutions. Different sets of initial parameters can provide reasonable fits to the data, if the initial compositions and degrees of crystallization are adjusted accordingly. Second, the fits are imperfect for Re vs. Pt, given that several irons plot near the solid and liquid tracks, but resolvably outside of their bounds. This may be due to the use of inaccurate D values, scatter in the Pt abundances caused by some process other than fractional crystallization, inaccurate classification of some irons, or a combination thereof. Third, the degree of fractional crystallization at which certain meteorites are modeled to have crystallized (i.e., their position relative to the solid and liquid tracks and mixing lines) is not always consistent when calculated using different element pairs (e.g., Magura, Morasko). This can be attributed to significant scatter in the relative HSE abundances in the MG, as evidenced by the disordered nature of the HSE patterns in the MG, in which crossing of the HSE patterns is common. Finally, if as discussed above, the Morasko cluster of irons (Morasko/Seeläsgen, Burgavli, and Sarepta) is distinct from the MG (e.g., Pilski et al., 2013), the low end of the compositional range of the MG can more easily be accounted for by CS alone. However, the CS-FC model, including the initial HSE and light element abundances of the CS portion of the model, are still required to account for the meteorites with high HSE abundances (e.g., Osseo, Magura, and Cranbourne).

We conclude that the compositional range of the MG, the abundant silicate inclusions, and evidence for mixing can be best accounted for by the combined CS-FC model. The importance of the CS model in the crystallization of the MG is in general agreement with that of

Wasson and Kallemeyn (2002), but the additional fractional crystallization of some portion of the liquid endmember can better account for the lower Re, Os, and Ir concentrations, as well as the scatter in plots of Re vs. Ru and Pt. The involvement of fractional crystallization is also in agreement with the observation that the more fractionated HSE patterns for some MG samples share similarities with those of some magmatic IVA iron meteorites.

5.3.3 The physical scenario for the CS-FC model

The scenario we envision for the CS-FC process to form the MG is as follows: an initial metallic melt, produced either by impact-melting (e.g., Wasson and Kallemeyn, 2002) or during the partial differentiation of an internally heated body (e.g., Benedix et al., 2000), crystallized at equilibrium until ~15% solidification was achieved. At this point, the crystals and liquid were separated by a process, such as filter pressing (Kracher, 1985; Ruzicka et al., 2006), defining the liquid and solid endmember compositions. These endmember components were then re-mixed in varying proportions, re-equilibrated, and the liquid metal crystallized. Re-mixing of the endmember components could be facilitated either by additional impacts or by density-driven overturn of the melt pool(s). The range of the resulting compositions is represented by the iron meteorites that fall on the initial two-component mixing lines (e.g., Landes), and is consistent with mixed compositions of up to 60% solid (thick black line in **Fig. 7**). We speculate that as much as 40% of the remaining liquid endmember was available for subsequent fractional crystallization. Our preferred model suggests that ~30% fractional crystallization of the remaining liquid can account for the samples with lower Ir abundances, and the scatter in plots of Re vs. Ru and Pt (4 to 8 of 23 MG iron meteorites studied here).

The stage at which the shift from equilibrium to fractional crystallization occurs is when some of the liquid endmember is no longer successfully mixed with the rest of the system. This could result from the downward migration of the liquid in the parent body to depths where it was hotter, before mixing of the liquid and solid metal and silicates could occur, or before crystallization could freeze-in silicates. In either case, the largely unmixed liquid endmember portion would have few or no entrained silicates and Fe-Ni crystals. Therefore, the fractionally crystallized samples (i.e., those with low Ir and/or high Pt and Ru) would be expected to contain fewer silicates, unless they were subsequently mixed-in by additional impacts. This is consistent with our survey of silicate occurrences in the MG (e.g., Buchwald, 1975; **Table S2**).

5.4. Trace element modeling of the subgroups

The sLM subgroup is characterized by a moderate range in HSE compositions and broadly linear trace element trends on plots of Re vs. HSE, with r^2 values > 0.7 for all but Re vs. Pd (**Fig. 8**). When the CS model is applied to the sLM subgroup, the ranges of HSE concentrations and the slopes of the Re vs. HSE are generally reproduced using $F = 0.25$ and D values consistent with 14 wt. % S, 2.8 wt. % P, and 0.2 wt. % C (IP-sLM_{CS} – **Table 3**). The average HSE concentrations of the sLM subgroup are used as the initial composition of the melt. If Pitts and Woodbine are included in the sLM subgroup, lower F and higher total light element contents are required to account for these samples (e.g., $F = 0.12$; 22 wt. % S, 1 wt. % P, and 0.2 wt. % C). The HSE patterns of the sLM subgroup are unlike those expected from fractional crystallization, and the lack of systematic variation in the relative abundances of Re, Os, and Ir indicates that the chemical diversity in the sLM subgroup was not a result of fractional

crystallization. The CS model alone can best account for the range of HSE compositions in the sLM subgroup. The similarity in the HSE patterns between the sLM and sLH subgroups suggests that the sLH subgroup also crystallized by CS.

The relative abundances of HSE in the sLL subgroup are generally chondritic, with a narrow range in absolute HSE abundances (**Fig. 3**). The presence of silicate inclusions in most sLL iron meteorites (**Table S2**) argues against fractional crystallization, and the limited range in HSE concentrations of the sLL subgroup makes it difficult to assess the applicability of the CS model (**Fig. S7**). Qualitatively, the HSE patterns and absolute abundances are similar to those of H chondrite metal fractions (Horan et al., 2009). Therefore, the sLL subgroup most likely represents metal that was batch melted and extracted from a chondritic precursor, and was subsequently crystallized without significant crystal-liquid fractionation.

As noted earlier, it is questionable whether the nominal sHL and sHH subgroups constitute genetically significant groups, but the HSE patterns share some similarities with those of the later crystallized IIAB and IIIAB iron meteorites, suggesting that some of them may have resulted from fractional crystallization. If it is assumed that the irons in the individual subgroups are related, the Re-Os, Pt-Os, and Pd element systems show a larger range of compositions than can be explained by fractional crystallization with the initial light element parameters examined here for either subgroup (**Fig. S8**). Therefore, if these irons crystallized by fractional crystallization, as is plausible based on their highly fractionated HSE patterns, two or fewer meteorites from each subgroup are related. Further, silicates are present in the anomalous Sombrerete and in the substantially fractionated Mount Magnet, which may preclude an interpretation of fractional crystallization. Ruzicka et al. (2006) argued for a complex partial melting scenario to account for the unusual characteristics of Sombrerete, which may be

applicable to those irons with similar HSE patterns. Though not useful for determining crystallization mechanisms in these subgroups, the modeling here provides additional evidence that the iron meteorites within the sHL and sHH subgroups may not have crystallized from the same parental melt.

5.5. Initial melt compositions in the IAB complex

The HSE compositions of the initial melts calculated from the CS-FC model for the MG, the CS model for the sLM subgroup, and the average HSE concentrations for the sLL subgroup are shown in Fig. 9 and provided in **Table 5**. The relative abundances of the MG initial melt are most similar to the LL chondrite Chainpur (**Fig. 9**). This implies a chondritic precursor, which is in agreement with previous work (e.g., Benedix et al. 2000; Wasson and Kallemeyn, 2002). Due to the minor fractionation in the sLL subgroup, it can be assumed that HSE abundances in the initial melt in this subgroup were also chondritic. The initial melt composition of the sLM subgroup has both absolute and relative abundances of Re, Os, Ir, Ru, and Pt that are similar to bulk chondrites, but is strongly enriched in Pd. Similar to Pd, the Au abundances reported by Wasson and Kallemeyn (2002) are also enriched. The initial melt of the analogous sLH subgroup was likely similar. Therefore, if the sLM and sLH subgroups crystallized *via* the CS model, the precursor materials were subjected to some prior nebular or planetary processing to either raise the Pd and Au concentrations, or lower the other HSE concentrations. Because Pd and Au are more volatile than the other HSE, the characteristic HSE patterns in the initial sLM and sLH melts most likely resulted from a volatility-driven processes. The precursor material for these groups may have been a late-stage condensate, which would result in relative and absolute

refractory element abundances that were broadly chondritic, but higher abundances of the more volatile Pd and Au. Complementary HSE patterns, in which Pd is depleted in comparison to the chondritic relative abundances of the more refractory HSE, have been observed in early nebular condensates, such as Ca-Al rich inclusions (CAI) (Archer et al., 2014). Although we do not expect that the sLM and sLH precursor materials and CAIs were condensed from the same materials, the complementary patterns indicate that volatility-driven processes were important in their formation.

5.6. Formation of the IAB complex

One unresolved question relating to the IAB complex is what larger scale processes can account for the smaller scale crystallization processes that were modeled above. The mechanisms which were required to generate each of the subgroups, including the heat source and the nature of the mixing process, are discussed here.

The spectrum of MG metals can be accounted for by crystal segregation, coupled with fractional crystallization, wherein mixing processes were important. Three important characteristics of MG irons, in the order of shortest to longest timescales required to generate them are: 1) retention of silicates which, in the MG, are largely confined to the least fractionated irons, 2) evidence for fractional crystallization, and 3) slow sub-solidus metallographic cooling rates (e.g., Herpfer et al., 1994; Goldstein et al., 2013).

These conditions can potentially be met in a scenario similar to that proposed by Benedix et al. (2000) for the IAB complex as a whole. In such a model, the MG metal partially melted and segregated due to internal heating, and the metal slowly crystallized by fractional

crystallization in a core or large, sinking, or pooled diapirs. Over a short time period, at least one impact could have re-melted metal, mixed silicates, and allowed for swift crystallization of some portion of the metal; the products of which were finally reburied to cool slowly sub-solidus. However, for models requiring dominantly internal heating, such as the Benedix et al. (2000) and Kracher (1985) models, fractionally crystallized metal is expected to be more volumetrically significant than is observed for the MG. Further, more fractionated HSE patterns than are observed in the MG would also be expected to result from fractional crystallization of a S-rich core, as proposed by Kracher (1985). Because irons which may have resulted from more extensive fractional crystallization (some ungrouped, sHL, or sHH irons) are not from the same parental melt as the MG, these cannot be invoked to represent a missing fractionally crystallized reservoir. Moreover, evidence from the short-lived radiogenic Hf-W system suggests that the metal-silicate segregation age of the MG meteorite Campo del Cielo was ~5 Ma after the formation of CAIs (Schulz et al., 2012; calculated using the CAI initial $^{182}\text{W}/^{183}\text{W}$ from Kruijer et al., 2014a). This metal-silicate segregation age is younger than those of most magmatic iron meteorite groups and is beyond the limit of the ~ 4 Ma effective lifetime of ^{26}Al heating (e.g., Kruijer et al. 2014b).

Our preferred model for the formation of the MG invokes a dominantly external heat source, where initial melting of metal and mixing of silicates was coincident due to an impact event, followed by the processes discussed in section 5.3.3, including partial equilibrium crystallization, filter-pressing enabled separation of liquid and solid, and impact-induced or density-driven mixing of solid and liquid metal and silicates. Finally, fractional crystallization of the remaining, but volumetrically minor, portion of the liquid occurred after it descended into the warmer interior. We envision that the reduced viscosity of the descending melt, due to the lack of

suspended solids and higher temperatures, would slow solidification and would allow for fractional crystallization to occur.

The first irons to crystallize must have done so quickly to prevent the separation of metal and silicate, and the paucity of silicate inclusions in the fractionated MG irons indicates that the metallic liquid was separated from silicate material as crystallization progressed. The evidence for fractional crystallization and the relatively uniform, slow sub-solidus cooling rates suggest that the MG meteorites crystallized at greater depth within the parent body and/or were reburied. Therefore, the preferred MG formation model is best explained by impact heating because a dominant external heat source allows events to take place in order of short to long timescale, as would be expected in a relatively simple heating and cooling profile. Further, impact heating provides a mechanism for mixing silicates, liquid metal, and solid metal, and for subsequent reburial.

The presence of silicates and the slow metallographic cooling rates in the sLM and sLL subgroups, along with the observation that CS model can account for the HSE variations in the sLM subgroup and the limited fractionation in the sLL subgroup, indicate that mixing and burial processes were important. For the reasons outlined above for the MG, impact heating is also considered the most likely source of melting for both the sLM and sLL subgroups. We note, however, that the CS-FC model of the MG and the CS model of the sLM subgroup are applicable regardless of the original source of the melt.

Although impact heating is the preferred mechanism by which to generate the initial MG, sLM, and sLL melts, some degree of internal heating is likely. The size of the IAB parent body or bodies is not well constrained, but diameter estimates range from tens to hundreds of

751 kilometers (e.g., Kracher, 1985; Rasmussen, 1989; Benedix et al., 2000). If the IAB parent
752 body(ies) was large enough for heat retention to be effective, ^{26}Al and other short-lived
753 radionuclides could contribute to internal heating, though perhaps not to melting temperatures.
754 Studies of winonaites and silicate inclusions in IAB meteorites indicate that these meteorites
755 experienced heating corresponding to a minimum of grade 6 metamorphism in chondrites
756 (Takeda et al., 2000; Benedix et al., 2000). Because this evidence is from both winonaites and
757 IAB irons, this heating was likely on a large to global scale. Keil (1997) and Davison (2013),
758 however, concluded that impact heating is not efficient on global scales. For example, for a 100
759 km diameter, 20% porous body, Davison et al. (2013) concluded using Monte-Carlo simulations
760 that heating resulting from all the impacts that occurred over a 100 Ma period can raise the
761 global average temperature of a body by a maximum of only ~ 35 K without disrupting the body.
762 If a disruptive event occurred, the globally averaged temperature increase can be up to 1000 K.
763 The more likely low velocity, oblique impacts expected for planetary bodies early in Solar
764 System history probably resulted in smaller global temperature increases. These authors did not
765 account for heating from ^{26}Al and assumed an initial temperature of 300 K. The porosity used by
766 Davison (2013) is likely applicable to the initial porosity of the IAB parent body(ies), as is the
767 size, because 20% porosity is between the porosity of chondritic meteorites and that estimated
768 for most asteroids (Britt et al., 2002).

769 Therefore, it seems unavoidable that internal heating played some role, if only to warm
770 the parent body to metamorphic temperatures such that impact heating on local scales was more
771 efficient. Impacts on porous chondritic bodies are also conducive to producing significant
772 volumes of melt, especially locally (e.g., Davison et al., 2010). Thus, if impact heating was the
773 primary heat source, as suggested by the HSE abundances and crystallization modeling discussed

above, the MG, sLL, and sLM subgroups originated in locally impact-heated portions of a $\geq 20\%$ porous, large (> 100 km diameter) parent body(ies) where metamorphic temperatures were reached and retained by internal heating from ^{26}Al . Alternately, these subgroups may have originated from a disruptive impact or impacts, in which impact-heating is more efficient.

5.7. Diversity among IAB complex subgroups

Potential sources for the variability of IAB complex subgroups are considered here. The similarity of the sLL and MG silicates (e.g., Benedix et al., 2000), and the absolute abundances of many major and trace elements, implies that these meteorites were formed by similar processes from similar starting materials. Variable degrees of impact melting could lead to distinct initial compositions (Choi et al., 1995; Wasson and Kallemeyn, 2002). We postulate that the volume of melt associated with the MG was larger than the sLL subgroup, so that the larger volume of MG melt took longer to cool, and could migrate and fractionally crystallize, whereas the sLL melt simply froze in place. The new results, however, provide no information regarding whether or not the MG and sLL subgroup were generated on the same parent body.

Similarly to the sLL subgroup and MG, the sLM and sLH subgroups may have originated in multiple impact-generated melt pools. In the case of these subgroups, it is likely they formed on one or more parent bodies which were not shared with the MG. This is because the initial sLM and sLH melts were likely non-chondritic (**Fig. 9**), and the light element abundances were potentially higher than in the MG.

The siderophile element evidence and modeling outlined above is most consistent with the formation of most IAB subgroups in separate impact-generated melts, in agreement with

Wasson et al. (1980), Wasson and Kallemeyn (2002), and Tomkins et al. (2013). Evidence for different parental melts for the IAB complex subgroups has several implications for the possibility that the IAB complex was sourced from multiple parent bodies. Because the IAB subgroups are so diverse in terms of their relative and absolute siderophile element abundances, and may have crystallized in distinct light element regimes, it is unlikely that all IAB complex subgroups originated on the same asteroid-sized parent body.

This possibility can be tested using isotopic genetic tracers (e.g., O, Mo, Ru – Clayton and Mayeda, 1996; Dauphas et al., 2002; Greenwood et al., 2006; Franchi, 2008; Chen et al., 2010; Burkhardt et al., 2011; Fischer-Gödde et al., 2015). Isotopic data in these and other genetically useful isotope systems are limited to one or two meteorites for most IAB subgroups excluding the MG, sLL subgroup and ungrouped IAB irons. For example, Carlton (sLM) and Dayton (sLH) have similar a Mo isotopic composition and $\Delta^{17}\text{O}$, respectively, to other IAB irons (Clayton and Mayeda, 1996; Burkhardt et al., 2011). Sombroete and Magnesia (sHL) have a distinct $\Delta^{17}\text{O}$ and Mo isotopic composition, respectively, from other IAB irons (Clayton and Mayeda, 1996; Dauphas et al., 2002). Thus, it is likely that some sHL subgroup irons are genetically unrelated to the MG, whereas the existing isotopic data for sLL, sLM and sLH meteorites are permissive, but not confirmation, of a common parent body with the MG. No genetic isotope data are published for sHH subgroup irons.

6. CONCLUSIONS

Multiple formational modes are represented in the IAB complex. Moreover, the models used to constrain the formational processes require different initial compositions and variable

818 light element contents to account for the observed HSE ranges of each subgroup. Broadly, this
819 conclusion points to a diversity of compositions and crystallization processes active on a single
820 parent body or a number of similar parent bodies. The HSE abundances reported here,
821 particularly Pd abundances, support the division of the complex into the subgroups suggested by
822 Wasson and Kallemeyn (2002). The variations in the relative and absolute abundances of HSE
823 among the subgroups are consistent with the interpretation that the subgroups originated in
824 different parental melts. Therefore, for future studies to draw conclusions about the IAB
825 complex, IAB meteorites should be considered in the context of their subgroup rather than with
826 the complex as a whole.

827 The Re-Os isotopic systematics of the complex indicate that the HSE abundances of most
828 IAB irons reflect those at the time of crystallization, although minor open system behavior,
829 which likely occurred ≥ 40 Ma after crystallization, is evident in the sLM, sLH, sHL, and sHH
830 subgroups. The conclusion that the IAB subgroups originated in separate parental melts in
831 different locations is supported by the observation that only some subgroups show open system
832 behavior.

833 The MG of the complex reflects metal that was crystallized by crystal segregation,
834 coupled with fractional crystallization. The initial light element abundances of the system were
835 probably relatively high, with the best-fit crystallization models based on the assumption of ~ 12
836 wt. % total S, P, and C. The preferred scenario for the formation of the MG is that impact-
837 generated melting initially produced metallic melt, into which silicates were mixed. The melt
838 cooled and mixed quickly at first, freezing silicates in place, but some portion of the melt
839 migrated to warmer depths in the body, where the melt entered a fractional crystallization

regime. The scarcity of observed silicates in the more chemically evolved MG irons is consistent with this interpretation.

The most likely formation scenario of the sLL subgroup is an impact-generated melt origin, similar to that of the MG, but the minimally fractionated HSE patterns point to important differences between the volume of the melt, the nature of the impact, or a combination thereof. The relative abundances of the HSE calculated for the initial melts of both the MG and sLL subgroup are roughly chondritic. This suggests that their parent body or bodies had chondritic bulk compositions with limited nebular or planetary processing of their precursor materials prior to metal-silicate segregation.

In the sLM subgroup, the distinctive HSE patterns and the broadly linear distributions of the relative HSE abundances suggest that it represents irons crystallized by crystal segregation. The sLM subgroup may have crystallized from a system with total initial light element contents significantly higher than those of the MG, between ~17-23 wt. %. The sHL subgroup probably formed in a similar manner to the sLM subgroup, based on the comparable HSE patterns. An impact origin is also considered likely in these subgroups; however, the non-chondritic HSE pattern of the initial melt and the higher initial light element abundances of the sLM system indicate that the precursor materials were unrelated to the MG. The enriched Pd and Au relative to Re, Os, Ir, Ru, and Pt abundances in the sLM initial melt suggest that the sLM precursor materials may represent late-stage condensates. The irons in the sHL and sHH subgroups may have been fractionally crystallized and, in the case of the sHL subgroup, formed on a separate parent body from the MG.

Acknowledgements: We gratefully acknowledge the Smithsonian Institute National Museum of Natural History for providing the samples used in this study. Funding for this work was provided by NASA grant NNX13AF83G. We thank an anonymous reviewer, reviewers John Wasson, Ed Scott, and Nancy Chabot, and Associate Editor Munir Humayun for their highly constructive comments. We also thank Igor Puchtel and Richard Ash for lab and analytical guidance; Lynnette Pitcher for conducting the analyses for some of the ungrouped IAB iron meteorites; and Greg Archer, Bill Bottke, and Tim McCoy for helpful discussions.

References:

- Albarède F., Bouchet R. A., and Blichert-Toft J. (2013) Siderophile elements in IVA irons and the compaction of their parent asteroidal core. *Earth Planet Sc. Lett.* **362**, 122-129.
- Archer G. J., Ash R. D., Bullock E. S., and Walker R. J. (2014) Highly siderophile elements and ^{187}Re - ^{187}Os isotopic systematics of the Allende meteorite: Evidence for primary nebular processes and late-stage alteration. *Geochim. Cosmochim. Acta* **131**, 402-414.
- Benedix G.K., McCoy T.J., Keil K., and Love S.G. (2000) A petrologic study of the IAB iron meteorites: Constraints on the formation of the IAB-Winonaite parent body. *Meteorit. Planet. Sci.* **35**, 1127-1141.
- Bild R.W. (1977) Silicate inclusions in group IAB irons and a relation to the anomalous stones Winona and Mt. Morris (Wis). *Geochim. Cosmochim. Acta* **41**, 1439-1456.
- Birck J.-L., Roy-Barman M., and Capmas F. (1997) Re-Os isotopic measurements at the femtomole level in natural samples. *Geostand. Newslett.* **20**, 9-27.
- Bogard D.D., Garrison D.H., and Takeda H. (2005) Ar-Ar and I-Xe ages and the thermal history of IAB meteorites. *Meteorit. Planet. Sci.* **40(2)**, 207-224.
- Britt D.T., Yeomans D., Housen K., Consolmagno G. (2002) Asteroid density, porosity, and structure. In: *Asteroids III* (eds. Bottke, W.F., Jr., Cellino, A., Paolicchi, P., Binzel, R.P.), pp. 485–500. University of Arizona Press, Tucson.
- Buchwald V.F. (1975) *Handbook of Iron Meteorites*. University of California Press. 1418 pp.
- Bunch T. E., Keil K., and Huss G. I. (1972) The Landes meteorite. *Meteoritics* **7**, 31-38.
- Burkhardt C., Kleine T., Oberli F., Pack A., Bourdon B., and Wieler R. (2011). Molybdenum isotope anomalies in meteorites: Constraints on solar nebula evolution and origin of the Earth. *Earth Planet. Sc. Lett.* **312**, 390-400.
- Chabot N.L. and Jones J.H. (2003) The parameterization of solid metal-liquid metal partitioning of siderophile elements. *Meteorit. Planet. Sci.* **38**, 1425-1436.
- Chabot N.L., Wollack E.A., McDonough W.F., and Ash R.D. (2014) The effect of light elements in metallic liquids on partitioning behavior. *Lunar Planet. Sci.* XLV. Lunar Planet. Inst., Houston. #1165 (abstr.).
- Chen J. H., Papanastassiou D. A., and Wasserburg G. J. (2010) Ruthenium endemic isotope effects in chondrites and differentiated meteorites. *Geochim. Cosmochim. Acta* **74**, 3851-3862.
- Choi B. G., Ouyang X. W., and Wasson J. T. (1995) Classification and Origin of IAB and IIICD Iron-Meteorites. *Geochim. Cosmochim. Acta* **59**, 593-612.

- Clayton R. N. and Mayeda T. K. (1996) Oxygen isotope studies of achondrites. *Geochim. Cosmochim. Acta* **60**, 1999-2017.
- Cohen A. S. and Waters F. J. (1996) Separation of osmium from geological materials by solvent extraction for analysis by thermal ionization mass spectrometry. *Anal. Chim. Acta* **332**, 269-275.
- Cook D. L., Walker R. J., Horan M. F., Wasson J. T., and Morgan J. W. (2004) Pt-Re-Os systematics of group IIAB and IIIAB iron meteorites. *Geochim. Cosmochim. Acta* **68**, 1413-1431.
- Crocket J. H. (1972) Some aspects of the geochemistry of Ru, Os, Ir, and Pt in iron meteorites. *Geochim. Cosmochim. Acta* **36**, 517-535.
- Dauphas N., Marty B., and Reisberg L. (2002) Molybdenum evidence for inherited planetary scale isotope heterogeneity of the protosolar nebula. *Astrophys. J.* **565**, 640-644.
- Davison T. M., Collins G. S., and Ciesla F. J. (2010) Numerical modelling of heating in porous planetesimal collisions. *Icarus* **208**, 468-481.
- Davison T. M., O'Brien D. P., Ciesla F. J., and Collins G. S. (2013) The early impact histories of meteorite parent bodies. *Meteorit. Planet. Sci.* **48**, 1894-1918.
- Fischer-Gödde M., Burkhardt C., Kruijer T. S., and Kleine T. (2015) Ru isotope heterogeneity in the solar protoplanetary disk. *Geochim. Cosmochim. Acta* **168**, 151-171.
- Fish R.A., Goles G.G., and Anders E. (1960) The record in the meteorites. III. on the development of meteorites in asteroidal bodies. *Astrophys. J.* **132**, 243.
- Franchi I. A. (2008) Oxygen isotopes in asteroidal materials. *Rev. Mineral. Geochem.* **68**, 345-397.
- Goldberg E., Uchiyama A., and Brown H. (1951) The distribution of nickel, cobalt, gallium, palladium and gold in iron meteorites. *Geochim. Cosmochim. Acta* **2**, 1-25.
- Goldstein J. I. and Short J.M. (1967) The iron meteorites, their thermal history and parent bodies. *Geochim. Cosmochim. Acta* **31**, 1733-1770.
- Goldstein J. I., Scott E. R. D., and Chabot N. L. (2009) Iron meteorites: Crystallization, thermal history, parent bodies, and origin. *Chem. Erde-Geochem.* **69**, 293-325.
- Goldstein J. I., Scott E. R. D., Winfield T., and Yang J. (2013) Thermal histories of group IAB and related iron meteorites and comparison with other groups of irons and stony iron meteorites. *Lunar Planet. Sci. XLIV*. Lunar Planet. Inst., Houston. #1394 (abstr.)

- Greenwood R. C., Franchi I. A., Jambon A., Barrat J. A., and Burbine T. H. (2006) Oxygen isotope variation in stony-iron meteorites. *Science* **313**, 1763-1765.
- Herpfer M. A., Larimer J. W., and Goldstein J. I. (1994) A comparison of metallographic cooling rate methods used in meteorites. *Geochim. Cosmochim. Acta* **58**, 1353-1365.
- Hirata T. and Masuda A. (1992) Rhenium and osmium systematics on iron and stony iron meteorites. *Meteoritics* **27**, 568-575.
- Hirata T. and Nesbitt R. W. (1997) Distribution of platinum group elements and rhenium between metallic phases of iron meteorites. *Earth Planet. Sc. Lett.* **147**, 11-24.
- Hoashi M., Brooks R.R., and Reeves R.D. (1993) Palladium, platinum, and ruthenium in iron meteorites and their taxonomic significance. *Chem. Geol.* **106**, 207-218.
- Horan M. F., Smoliar M. I., and Walker R. J. (1998) ^{182}W and ^{187}Re - ^{187}Os systematics of iron meteorites: Chronology for melting, differentiation, and crystallization in asteroids. *Geochim. Cosmochim. Acta* **62**, 545-554.
- Horan M. F., Walker R. J., Morgan J. W., Grossman J. N., and Rubin A. E. (2003) Highly siderophile elements in chondrites. *Chem. Geol.* **196**, 5-20.
- Horan M. F., Alexander C. M. O'D., and Walker R. J. (2009) Highly siderophile element evidence for early Solar System processes in components from ordinary chondrites. *Geochim. Cosmochim. Acta* **73**, 6984-6997.
- Jones J.H. and Drake M.J. (1983) Experimental investigations of trace element fractionation in iron meteorites. II: The influence of sulfur. *Geochim. Cosmochim. Acta* **47**, 1199-1209.
- Jones J.H. and Malvin D.J. (1990) A nonmetal interaction model for the segregation of trace metals during solidification of Fe-Ni-S, Fe-Ni-P, and Fe-Ni-S-P alloys. *Metall. Mater. Trans. B* **21B**, 697-706.
- Kelly W. R. and Larimer J. W. (1977) Chemical fractionations in meteorites – VIII. Iron meteorites and the cosmochemical history of the metal phase. *Geochim. Cosmochim. Acta* **41**, 93-111.
- Keil K., Stöffler D., Love S. G., and Scott E. R. D (1997) Constraints on the role of impact heating and melting in asteroids. *Meteorit. Planet. Sci.* **32**, 349-363.
- Kleine T., Mezger K., Münker C., Palme H., Bischoff A. (2004) ^{182}Hf - ^{182}W isotope systematics of chondrites, eucrites, and Martian meteorites: Chronology of core formation and mantle differentiation in Vesta and Mars, *Geochim. Cosmochim. Acta* **68** 2935–2946.
- Kracher A. (1985) The evolution of partially differentiated planetesimals: Evidence from iron meteorite groups IAB and IIICD. *J. Geophys. Res.* **90** (S02), C689-C698.

- Kracher A., Willis J., and Wasson J. T. (1980) Chemical classification of iron meteorites-IX. A new group (IIF), revision of IAB and IIICD, and data on 57 additional irons. *Geochim. Cosmochim. Acta* **44**, 773-787.
- Kruijer T. S., Kleine T., Fischer-Gödde M., Burkhardt C., and Wieler R. (2014) Nucleosynthetic W isotope anomalies and the Hf-W chronometry of Ca–Al-rich inclusions. *Earth Planet. Sc. Lett.* **403**, 317-327.
- Kruijer T. S., Touboul M., Fischer-Gödde M., Bermingham K. R., Walker R. J., and Kleine T. (2014) Protracted core formation and rapid accretion of protoplanets. *Science* **344**, 1150-1154.
- Lodders K. (2003) Solar System abundances and condensation temperatures of the elements. *Astrophys. J.* **59**, 1220-1247.
- Lovering J. F. (1957) Differentiation in the iron-nickel core of a parent meteorite body. *Geochim. Cosmochim. Acta* **12**, 238-252.
- Lovering J. F., Nichiporuk W., Chodos A., and Brown H. (1957) The distribution of gallium, germanium, cobalt, chromium, and copper in iron and stony-iron meteorites in relation to nickel content and structure. *Geochim. Cosmochim. Acta* **11**, 263-278.
- Ludwig K. R. (2003) User's Manual for Isoplot 3.00. Berkeley Geochronology Center Special Publication No. 4, Berkeley, CA, 70pp.
- McCoy T. J., Keil K., Scott E. R. D, and Haack H. (1993) Genesis of the IIICD iron meteorites: Evidence from silicate-bearing inclusions. *Meteoritics* **28**, 552-560.
- McCoy T. J., Ehlmann A. J., Benedix G. K., Keil K., and Wasson J. T. (1996) The Lueders, Texas, IAB iron meteorite with silicate inclusions. *Meteorit. Planet. Sci.* **31**, 419-422.
- McCoy T. J., Walker R. J., Goldstein J. I., Yang J., McDonough W. F., Rumble D., Chabot N. L., Ash R. D., Corrigan C. M., Michael J. R., and Kotula P. G. (2011) Group IVA irons: New constraints on the crystallization and cooling history of an asteroidal core with a complex history. *Geochim. Cosmochim. Acta* **75**, 6821-6843.
- Moore C. B., Lewis C. F., and Nava D. (1969) Superior analyses of iron meteorites. In *Meteorite Research* (ed. P. M. Millman), pp. 738-748. Reidel, Dodrecht, Holland.
- Pernicka E. and Wasson J.T. (1987) Ru, Re, Os, Pt, and Au in iron meteorites. *Geochim. Cosmochim. Acta* **51**, 1717-1726.
- Pilski A. S., Wasson J. T., Muszyński A., Kryza R., Karwowski Ł., and Nowak M. (2013) Low-Ir IAB irons from Morasko and other locations in central Europe: One fall, possibly distinct from IAB-MG. *Meteorit. Planet. Sci.* **48**, 2531-2541.

- Rasmussen K. L. (1989) Cooling rates and parent bodies of iron meteorites from group III CD, IAB, and IVB. *Phys. Scripta* **39**, 410-416.
- Rasmussen K. L., Malvin D. J., Buchwald V. F., and Wasson J. T., (1984) Compositional trends and cooling rates of group IVB iron meteorites. *Geochim. Cosmochim. Acta* **48**, 805-813.
- Rehkämper M., Halliday A. N., Barfod D., Godfrey F. J., and Dawson B. J. (1997) Platinum-group element abundance patterns in different mantle environments. *Science* **278**, 1595.
- Ruzicka A. (2014) Silicate-bearing iron meteorites and their implications for the evolution of asteroidal parent bodies. *Chem. Erde-Geochem.* **74**, 3-48.
- Ruzicka A. and Hutson M. (2010) Comparative petrology of silicates in the Udei Station (IAB) and Miles (IIE) iron meteorites: Implications for the origin of silicate-bearing iron meteorites. *Geochim. Cosmochim. Acta* **74**, 394-433.
- Ruzicka A., Hutson M., and Floss C. (2006) Petrology of silicate inclusions in the Sombrerete ungrouped iron meteorite: Implications for the origins of IIE-type silicate-bearing irons. *Meteorit. Planet. Sci.* **41**, 1797-1831.
- Schulz T., Upadhyay D., Münker C., and Mezger K. (2012) Formation and exposure history of non-magmatic iron meteorites and winonaite: Clues from Sm and W isotopes. *Geochim. Cosmochim. Acta* **85**, 200-212.
- Scott E. R. D. (1972) Chemical Fractionation in Iron-Meteorites and Its Interpretation. *Geochim. Cosmochim. Acta* **36**, 1205-1236.
- Scott E. R. D. and Wasson J. T. (1975) Classification and Properties of Iron-Meteorites. *Rev. Geophys.* **13**, 527-546.
- Scott E. R. D., Wasson J. T., and Buchwald V. F. (1973) The chemical classification of iron meteorites – VII. A reinvestigation of irons with Ge concentrations between 25 and 80 ppm. *Geochim. Cosmochim. Acta* **37**, 1957-1983.
- Shen J. J., Papanastassiou D. A., and Wasserburg G. J. (1996) Precise Re-Os determinations and systematics in iron meteorites. *Geochim. Cosmochim. Acta* **60**, 2887-2900.
- Smales A. A., Mapper D., and Fouche K. F. (1967) The distribution of some trace elements in iron meteorites as determined by neutron activation. *Geochim. Cosmochim. Acta* **31**, 673-720.
- Smoliar M. I., Walker R. J., and Morgan J. W. (1996) Re-Os ages of group IIA, IIIA, IVA, and IVB iron meteorites. *Science* **271**, 1099-1102.

- Takeda H., Bogard D.D., Mittlefehldt D.W., and Garrison D.H. (2000) Mineralogy, petrology, chemistry, and ^{39}Ar – ^{40}Ar and exposure ages of the Caddo County IAB iron: Evidence for early partial melt segregation of a gabbro area rich in plagioclase–diopside. *Geochim. Cosmochim. Acta* **64**, 1311-1327.
- Theis K.J., Schönbächler M., Benedix G.K., Rehkämper M., Andreasen R., and Davies C. (2013) Palladium-silver chronology of IAB iron meteorites. *Earth Planet. Sc. Lett.* **361**, 402-411.
- Tomkins A.G., Mare E.R., and Reveschi M. (2013) Fe-carbide and Fe-sulfide liquid immiscibility in IAB meteorite, Campo del Cielo: Implications for iron meteorite chemistry and planetesimal core compositions. *Geochim. Cosmochim. Acta* **117**, 80-98.
- Vogel N. and Renne P.R. (2008) ^{40}Ar – ^{39}Ar dating of plagioclase grain size separates from silicate inclusions in IAB iron meteorites and implications for the thermochronological evolution of the IAB parent body. *Geochim. Cosmochim. Acta* **72**, 1231-1255.
- Walker R. J., Horan M. F., Morgan J. W., Becker H., Grossman J. N., and Rubin A. E. (2002) Comparative ^{187}Re – ^{187}Os systematics of chondrites: Implications regarding early Solar System processes. *Geochim. Cosmochim. Acta* **66**, 4187-4201.
- Walker R. J., McDonough W. F., Honesto J., Chabot N. L., McCoy T. J., Ash R. D., and Bellucci J.J. (2008) Modeling fractional crystallization of group IVB iron meteorites. *Geochim. Cosmochim. Acta* **72**, 2198-2216.
- Wasson J. T. (1967) The chemical classification of iron meteorites: I. A study of iron meteorites with low concentrations of gallium and germanium. *Geochim. Cosmochim. Acta* **31**, 161-180.
- Wasson J. T. (1969) The chemical classification of iron meteorites – III. Hexahedrites and other irons with germanium concentrations between 80 and 200 ppm. *Geochim. Cosmochim. Acta* **33**, 859-876.
- Wasson J. T. (1970) The chemical classification of iron meteorites – IV. Irons with Ge concentrations greater than 190 ppm and other meteorites associated with group I. *Icarus* **12**, 407-423.
- Wasson J. T. (1974) *Meteorites: Classification and Properties*. Springer-Verlag.
- Wasson J. T. (1999) Trapped melt in IIIAB irons; solid/liquid elemental partitioning during the fractionation of the IIIAB magma. *Geochim. Cosmochim. Acta* **63**, 2875-2889.
- Wasson J. T. (2011) Relationship between iron-meteorite composition and size: Compositional distribution of irons from North Africa. *Geochim. Cosmochim. Acta* **75**, 1757-1772.

1143 Wasson J. T. and Kallemeyn G. W. (2002) The IAB iron-meteorite complex: A group, five
 1144 subgroups, numerous grouplets, closely related, mainly formed by crystal segregation in
 1145 rapidly cooling melts. *Geochim. Cosmochim. Acta* **66**, 2445-2473.

1146 Wasson J. T. and Richardson J. W. (2001) Fractionation trends among IVA iron meteorites:
 1147 Contrasts with IIIAB trends. *Geochim. Cosmochim. Acta* **65**, 951-970.

1148 Wasson J. T. and Schaudy R. (1971) The chemical classification of iron meteorites – V. Groups
 1149 IIC and IID and other irons with germanium concentrations between 1 and 25 ppm.
 1150 *Icarus* **14**, 59-70.

1151 Wasson J. T., Willis J., Wai C. M., and Kracher A. (1980) Origin of iron meteorite groups IAB
 1152 and IIICD. *Z. Naturforsch.* **35**, 781-795.

1153 Wasson J. T., Huber H., and Malvin D. J. (2007) Formation of IIAB iron meteorites. *Geochim.*
 1154 *Cosmochim. Acta* **71**, 760-781.

1155

1156 Willis J. (1981) Antimony in iron meteorites. *Earth Planet. Sc. Lett.* **53**, 1-10.

1157

1158 Willis J. and Goldstein J. I. (1982) The effects of C, P, and S on trace element partitioning during
 1159 solidification in Fe-Ni alloys. *Proc. 13th Lunar and Planet. Sci. Conf., J. Geophys. Res.*
 1160 **87**, A435-A445.

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2 **Table 1.** Highly siderophile element abundances and Re-Os isotope data. Samples are listed in
3 decreasing Re concentration. Sample weights are given in grams. [†]Ni, in wt. %, data are from
4 Wasson and Kallemeyn (2002). All other concentrations are reported as ng/g.

Sample	Wt.	Ni [†]	Re	Os	Ir	Ru	Pt	Pd	¹⁸⁷ Re/ ¹⁸⁸ Os	¹⁸⁷ Os/ ¹⁸⁸ Os	Δ _{Os} ^b
IAB main group											
Landes ^a	0.025	6.55	354.8	4368	4293	9805	10500	3581	0.3912	0.12627	-0.3
Seligman	0.195	6.67	308.7	3891	3293	7754	8886	3502	0.3822	0.12563	0.5
Yenberrie	0.108	6.82	290.6	3926	3463	5536	6359	3647	0.3565	0.12365	1.0
Ballinger	0.011	6.78	279.9	3153	3137	6782	7179	3402	0.4278	0.12905	-1.4
Duplicate	0.039		274.1	3138	3111	6341	7352	3382	0.4209	0.12868	0.3
Cranbourne	0.074	6.82	275.0	3670	3614	9662	11380	3398	0.3608	0.12377	-1.2
Campo del Cielo ^a	0.053	6.68	274.1	3632	3308	6725	8213	3403	0.3635	0.12418	0.8
New Leipzig	0.097	6.68	272.4	2723	2904	8626	9898	3527	0.4824	0.13353	0.1
Magura ^a	0.049	6.44	269.2	3743	3648	9892	11430	3254	0.3464	0.12270	-0.4
Canyon Diablo ^a	0.071	6.93	231.7	2780	2560	5579	6959	3494	0.4016	0.12723	1.1
Odessa ^a	0.016	7.19	224.3	2501	2572	5392	6749	3795	0.4321	0.12931	-2.3
Jenkins ^a	0.129	6.92	211.6	2327	2207	5403	6772	3585	0.4382	0.13008	0.6
ALHA 77283	0.065	7.28	210.2	1891	2327	5687	6867	4526	0.5362	0.13777	-0.1
Youndegin ^a	0.092	6.83	202.4	2134	2217	6354	6960	4010	0.4572	0.13156	0.4
Smithville ^a	0.033	6.92	191.4	1884	2028	6096	6422	3734	0.4898	0.13415	0.5
Seymour	0.100	6.78	166.5	1673	1807	5636	6383	3627	0.4799	0.13340	0.7
Bogou ^a	0.091	7.33	158.9	1219	1747	5178	5999	3754	0.6291	0.14510	-0.3
Burgavali	0.078	6.64	132.1	759.2	1217	9283	10120	3305	0.8422	0.16195	-0.5
Morasko	0.084	6.76	106.8	799.0	1141	9612	10260	3306	0.6456	0.14649	0.5
Seeläsgen	0.040	6.59	104.3	863.4	1143	9643	10340	3314	0.5831	0.14148	0.0
Sarepta	0.122	6.59	100.7	845.1	1131	9372	10190	3287	0.5748	0.14099	1.5
Duplicate	0.140		103.3	849.2	1136	9995	10270	3432	0.5873	0.14183	0.1
Hope ^a	0.203	7.06	54.26	288.9	753.1	6584	5893	3449	0.9094	0.16727	-0.6
IAB sLL (low Au, low Ni) subgroup											
Pine River ^a	0.261	8.04	268.1	3515	3085	3812	4925	4286	0.3674	0.12451	1.0
Shrewsbury ^a	0.113	8.53	226.5	3588	2729	3384	4840	4539	0.3038	0.11949	1.1
Duplicate	0.035		226.6	3641	2717	4254	4897	4413	0.2995	0.11912	0.8
Comanche ^a	0.071	8.24	226.4	2698	2614	4577	5738	4499	0.4043	0.12740	0.7
Bahjoi ^a	0.140	7.94	220.2	2598	2429	4659	5739	4370	0.4084	0.12771	0.5
Toluca ^a	0.141	8.02	215.9	2232	1962	4351	5426	4283	0.4663	0.13228	0.4
Deport	0.124	8.05	215.8	2587	2429	4633	5613	4389	0.4018	0.12721	0.7
Balfour Downs ^a	0.098	8.38	212.0	2422	2302	4148	5613	4809	0.4218	0.12870	-0.2
Bischtübe ^a	0.236	7.78	201.6	2432	2326	3554	4870	3932	0.3994	0.12639	-5.5
Duplicate	0.117		204.8	2425	2338	3822	4890	4257	0.4068	0.12764	1.1
Goose Lake ^a	0.045	8.24	188.2	2217	2139	4509	6094	4056	0.4092	0.12779	0.7
IAB sLM (low Au, medium Ni) subgroup											
Persimmon Creek ^a	0.194	13.79	80.73	730.6	840.2	1360	2026	5585	0.5330	0.13755	0.3
Duplicate	0.113		71.03	671.7	829.6	1314	1919	5184	0.5101	0.13577	0.6
Mungindi	0.197	11.86	47.86	678.4	570.3	426.2	610.7	5881	0.3397	0.12235	1.4
Edmonton (KY)	0.064	12.90	38.86	650.6	508.4	521.8	586.9	5769	0.2874	0.11817	1.0
Duplicate	0.284		38.74	649.7	502.0	516.1	584.4	5647	0.2869	0.11816	1.3
Maltahöhe ^a	0.198	11.40	16.42	121.4	165.0	641.0	626.2	5390	0.6533	0.14731	2.6
Anoka	0.309	11.80	13.84	260.2	167.9	256.2	224.5	5722	0.2558	0.11566	0.8
Duplicate	0.072		14.05	260.8	162.2	253.1	221.1	5847			
Carlton ^a	0.190	13.23	7.113	140.0	92.49	97.70	101.9	6689	0.2444	0.11494	2.7
Lamesa	0.268	13.28	4.582	39.78	48.14	142.6	138.6	6639	0.5559	0.14070	14
Duplicate	0.199		4.706	39.65	47.86	142.7	139.0	6632	0.5727	0.14120	5.3

(continued)

5

6 **Table 1.** (continued)

Sample	Wt.	Ni†	Re	Os	Ir	Ru	Pt	Pd	¹⁸⁷ Re/ ¹⁸⁸ Os	¹⁸⁷ Os/ ¹⁸⁸ Os	Δ _{Os} ^b
IAB sLH (low Au, high Ni) subgroup											
Tazewell	0.180	17.06	7.512	134.7	87.84	62.04	85.93	8687	0.2683	0.11692	3.6
Dayton ^a	0.252	17.10	2.594	41.82	31.37	49.13	48.66	8262	0.2985	0.12035	14
Duplicate	0.225		2.590	40.53	28.68	48.21	46.73	8194	0.3077	0.12036	6.8
Freda	0.278	23.21	0.6164	8.043	7.500	30.00	23.02	12210	0.3692	0.12662	21
IAB sHL (high Au, low Ni) subgroup											
Quarat al Hanish	0.150	12.75	86.75	943.4	957.6	2214	2671	6066	0.4432	0.13048	0.6
Chebarkol	0.238	9.17	8.632	63.31	102.9	1328	1711	4578	0.6587	0.14680	-6.7
Sombrerete ^a	0.264	9.84	6.099	61.87	73.80	537.6	468.2	5369	0.4753	0.13297	0.2
IAB sHH (high Au, high Ni) subgroup											
Kofa	0.231	18.39	8.218	72.38	99.22	2554	3671	9650	0.5478	0.13872	0.2
ALHA 80104	0.295	15.86	7.584	70.94	90.42	2597	3225	8641	0.5157	0.13669	5.4
Mount Magnet ^a	0.299	14.57	0.6252	5.650	6.176	142.6	59.90	9212	0.5340	0.13922	16
IAB ungrouped											
TIL 91725 ^a	0.131	8.04	285.3	3784	3244	5151	5934	3743	0.3631	0.12412	0.5
Caddo County ^a	0.260	9.42	254.0	2728	2515	4696	6058	4267	0.4488	0.13091	0.5
Zagora ^a	0.123	9.36	243.0	3026	2655	4520	5413	3981	0.3867	0.12607	1.3
Mertzona ^a	0.088	9.08	188.0	1701	2160	3959	7395	3540	0.5332	0.13760	0.6
EET 84300 ^a	0.093	10.11	183.4	2839	2161	1995	2768	4710	0.3109	0.12022	2.8
Duplicate	0.127		176.8	2769	2085	1876	2694	4094	0.3070	0.11982	2.0
Woodbine ^a	0.230	10.05	177.9	2062	1984	2389	3053	6181	0.4156	0.12834	1.2
Pitts ^a	0.158	12.81	117.9	1433	1244	1780	2447	6954	0.3962	0.12682	1.3
Duplicate	0.097		105.8	1423	1392	1724	2450	6486	0.3582	0.12405	3.7
Duplicate	0.111		114.1	1491	1373	1790	2346	6433	0.3685	0.12418	-3.2
Mundrabilla ^a	0.060	7.50	81.62	751.9	998.0	3232	3825	3925	0.5237	0.13754	7.5
Duplicate	0.078		89.40	785.5	971.6	3243	3627	3828	0.5491	0.13830	-5.0
Waterville	0.116	7.63	34.86	203.2	366.8	3278	3187	3874	0.8303	0.16061	-4.6
Duplicate	0.126		33.33	186.0	348.4	3395	2804	3908	0.8671	0.16255	-14
Georgetown ^a	0.109	9.03	17.33	88.47	190.4	2105	1750	3601	0.9488	0.16983	-6.1
IAB unclassified, classified here as MG											
Osseo	0.068	6.71	510.3	7313	5934	8769	10690	3368	0.3360	0.12204	1.2
Sardis	0.137	6.73	185.3	1760	2022	7588	8292	3424	0.5078	0.13551	-0.2
Duplicate	0.075		185.8	1765	2024	7546	8289	3421	0.5074	0.13558	0.9

^aIAB irons in which silicate inclusions have been observed by Buchwald (1975) and McCoy et al. (1993). A more detailed summary of silicate occurrences is provided in Table S2.

^bΔ_{Os} is calculated by, Δ_{Os} = 10⁴(¹⁸⁷Os/¹⁸⁸Os_{IAB} - (0.09532 + 0.07918 x ¹⁸⁷Re/¹⁸⁸Os_{IAB})), where 0.09532 is the initial ¹⁸⁷Os/¹⁸⁸Os and 0.07918 is the slope of the MG isochron (Table 2).

Table 2. ^{187}Re - ^{187}Os isochron slopes and initial $^{187}\text{Os}/^{188}\text{Os}$ ratios for each IAB subgroup and the IAB complex as a whole.

Group	n ^a	Age (Myr)	$^{187}\text{Os}/^{188}\text{Os}_i$	MSWD
MG	23	4574 ± 20	0.09532 ± 0.00017	4.2
sLL	10	4552 ± 23	0.09554 ± 0.00016	0.80
sLM	9	4640 ± 100	0.09513 ± 0.00068	60
sLH	4	5500 ± 2400	0.091 ± 0.013	134
sHL	3	4379 ± 35	0.09696 ± 0.00032	1.2
sHH	3	4000 ± 11000	0.10 ± 0.11	69
IAB complex	70	4533 ± 37	0.09576 ± 0.00029	55

^an is the number of samples, including duplicate analyses

21

22 **Table 3.** Initial parameters of the light
 23 element contents of metallic melts used
 24 in the crystallization models for the MG
 25 and sLM subgroups, respectively.

IP^a	S wt. %	P wt. %	C wt. %
MG _{CS}	5.0	1.0	0.2
MG _{CS-FC}	10	2.0	0.05
sLM _{CS}	14	2.8	0.2

26 ^aSubscripts refer to the model, where
 27 CS=crystal segregation and CS-FC=crystal
 28 segregation-fractional crystallization in this
 29 and subsequent tables.

30

Table 4. D values calculated as described in text and supplemental material for trace element modeling using the initial parameters in Table 3.

	IP-MG _{CS}	IP-MG _{CS-FC} (i)	IP-MG _{CS-FC} (50)	IP-sLM _{CS}
Ir	3.02	7.37	298	26.5
Re	2.89	6.82	194	19.7
Os	3.46	8.92	413	32.5
Ru	1.75	2.97	28.6	6.13
Pt	1.81	3.96	104	12.6
Pd	0.63	0.691	1.23	0.887

Subscripts refer to the model for which these D values were used. The degree of fractional crystallization are given in parentheses, where i denotes the initial D values for a given set of initial parameters, and 50 refers to the D values after 50% fractional crystallization. The modelling is not conducted beyond 50% fractional crystallization because, after that point, D values for some HSE grow exponentially to values that are not physically meaningful. D values are parameterized as described in text.

Table 5. Initial compositions for trace element modeling of the MG, sLM, and sLL subgroups. Concentrations are in ng/g. The solid and liquid endmember concentrations are calculated after 35%, 15%, and 25% crystallization for the MG_{CS} and MG_{CS-FC}, and sLM_{CS} models, respectively. The HSE abundances of the initial melts for each subgroup are taken as the mean of the HSE abundances as described in text.

	Re	Os	Ir	Ru	Pt	Pd
MG_{CS}						
MG mean (high Ir)	256	3032	2948	7012	8113	3612
Solid endmember	446	5636	5214	9719	11450	2611
Liquid endmember	154	1630	1728	5555	6316	4151
MG_{CS-FC}						
MG mean (low Pt)	226	2520	2514	6113	7086	3684
Solid endmember	821	10270	9471	14010	19420	2671
Liquid endmember	121	1152	1286	4719	4910	3863
sLM_{CS}						
sLM mean	30	374	341	492	616	5945
Solid endmember	104	1371	1227	1321	1992	5424
Liquid endmember	5	42	46	216	157	6118
sLL mean	220	2701	2446	4244	5433	4362

Fig. 1. Ni vs. Os for the IIAB, IIIAB, IVA, and IVB magmatic iron meteorite groups and the IAB complex. Fields represent the range of each group. Data symbols for the magmatic groups are omitted for clarity, but are shown for the IAB complex subgroups. The magmatic groups have more linear Ni vs. Os, and smaller ranges in Ni than the IAB complex as a whole. Os data for the IAB complex are from this study. Other Os data are from Cook et al. (2004), Walker et al. (2008), and McCoy et al. (2011), and Ni data are from Wasson (1974), Buchwald (1975), Rasmussen et al. (1984), Wasson, (1999), Wasson and Richardson (2001), and Wasson and Kallemeyn (2002).

13

14 **Fig. 2.** a) $^{187}\text{Re}/^{188}\text{Os}$ vs ΔOs (parts in 10,000 deviations from the MG Re-Os isochron, calculated
 15 using the slope and initial $^{187}\text{Os}/^{188}\text{Os}$ of the MG isochron – see Table 1 for details). b)
 16 $^{187}\text{Re}/^{188}\text{Os}$ vs. $^{187}\text{Os}/^{188}\text{Os}$ plot for the IAB complex, with the MG isochron for reference. The
 17 MG Re-Os system defines a primordial isochron, which most other IAB meteorites fall on or
 18 near. The sLM, sLH, sHL, and sHH subgroups are characterized by containing some irons which
 19 plot 5-20 ΔOs units off the MG isochron. Duplicates are included in the calculation of the
 20 isochron. MSWD (mean squared weighted deviation) is a measure of scatter about the
 21 regression.

Fig. 3. CI chondrite-normalized HSE abundances for each of the IAB complex subgroups.

Abundances are normalized to those of Orgueil (Horan et al., 2003). The HSE are plotted from left to right in order of decreasing 50% condensation temperature (Lodders, 2003). The HSE were determined by isotope dilution methods as described in text. Note the different scale in (a-b) relative to (c-f). The grey field in (a) represents the range of the IVA magmatic iron meteorite group, for reference (McCoy et al., 2011). The dashed and dotted lines represent Jamestown and Fuzzy Creek, respectively. Two unclassified samples, Osseo and Sardis, plot within the MG

30 field, or along trend, on all HSE variation diagrams (**Fig. 5**). These samples likely belong to the
31 MG, and so are plotted in (a) with blue diamond and yellow circle symbols. In (e-f), the IIIAB
32 magmatic iron meteorites Grant (dashed lines), Thurlow (dash-dotted line), and Chupaderos
33 (dotted line) are shown for comparison.

34

Fig. 4. CI chondrite-normalized HSE abundances for the IAB ungrouped iron meteorites. Some ungrouped IAB irons have HSE patterns which share similarities with those of a particular subgroup, as described in the text.

Fig. 5. a) Au vs. Pd Concentrations for the IAB subgroups and ungrouped IAB irons. b) Ni vs. Pd abundances. Pd abundances are from this study, whereas Au and Ni data are instrumental neutron activation analysis data from Wasson and Kallemeyn (2002). The subgroups of Wasson and Kallameyn (2002) can be distinguished on the basis of Pd vs. Au abundances and, to a lesser extent, Pd vs. Ni. In (b) Pd and Ni are correlated, demonstrating that the Pd vs. Au plot provides similar resolution between the subgroups as the Ni vs. Au plot from Wasson and Kallemeyn (2002).

Fig. 6. Crystal segregation model of Wasson and Kallemeyn (2002) applied to MG meteorites for a) Re vs. Ir, b) Re vs. Os, c) Re vs. Ru, d) Re vs. Pt, and e) Re vs. Pd. In (a-b) and (e), the CS model is in general agreement with the observed data, but it cannot account for low Re-Ir-Os samples. In (c-d) the scatter in the Re vs. Ru and Pt plots and the low Re samples cannot be accounted for with the CS model. Wasson and Kellemeyn (2002) suggested the low Re, Os, Ir,

58 samples may be accounted for by minor fractional crystallization (FC). The grey star represents a
59 mixture of 80% solid endmember, 20% liquid endmember.

60

Fig. 7. CS-FC model applied to MG iron meteorites using parameterized D values to calculate a) Re vs. Re/Os, b) Pt vs. Pt/Os, c) Re vs. Ru, d) Re vs. Pt, and e) Re vs. Pd. The thick black line represents the CS model and the red and blue lines correspond to the evolving solid and liquid compositions generated by the FC model. The dotted lines represent mixtures of coexisting solids and liquids at 0%, 10%, and 20% fractional crystallization. The white star corresponds to

67 the initial liquid composition, which is the composition of the liquid endmember generated by
68 the CS portion of the model. The black star corresponds to the initial solid composition. The grey
69 star represents a mixture of 60 % solid and 40% liquid. All or most IAB-MG iron meteorites plot
70 on or between the solid and liquid evolution tracks, or near the CS mixing line in (a-e). The error
71 bars for Pd, Ru, and Pt are approximately the size of the symbols.

72

Fig. 8. Crystal segregation model applied to the sLM subgroup, using D values consistent with IP-sLM. The CS model reproduces the ranges and slopes of Re vs. HSE. The grey star corresponds to the 80% solid, 20% liquid composition. The sLH irons and Pitts and Woodbine are plotted because the analogous HSE patterns to those of the sLM subgroup suggest that they formed in a similar manner to the sLM subgroup.

Fig. 9. Initial melt compositions of the MG, sLM, and sLL subgroups. The initial melt of the MG is the average HSE concentrations of MG samples, excluding low Ir and high Pt samples, as calculated for the CS-FC model. The sLM initial melt is the average HSE concentrations of all sLM samples, as calculated for the CS model. The sLL initial melt is also taken as the average of the HSE concentrations. The MG and sLL initial melts have broadly chondritic relative HSE abundances, and the MG initial melt is similar to the LL chondrite Chainpur (Horan et al., 2003). The sLM subgroup is enriched in Pd, relative to the chondritic abundances of the other HSE shown here. The HSE abundances of the calculated initial melts are given in **Table 5**.