

## NOTES ON THE EVOLUTION OF THE IIIAB/PALLASITE PARENT BODY

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1. **Introduction:** Although meteorites from any layer of a differentiated planetesimal provide information on parent body history, those samples which show the relationship between different layers are of special significance. It is therefore important to verify whether we have samples of both the core and the core/mantle interface of the same body, and to study their mutual relationship. The majority of pallasites, the so-called main group [1], is very probably related to IIIAB iron meteorites [2]. However, calculations concluding that the parent liquid of IIIAB irons after ~85% crystallization was injected into the overlying olivine layer [2] were based on constant partition coefficients [3], an assumption invalidated by recent experimental investigations [4-7], and the IIIAB-pallasite connection has been doubted [8].

In our view, element fractionations and mineralogy of IIIAB irons, main group pallasites, and possibly a quartet of ungrouped pallasites (Phillips County, Rawlinna, Springwater and Zaisho), can be understood by one consistent genetic scenario. Reconsideration of the crystallization history of iron meteorites reaffirms a single parent body for IIIAB's and main group pallasites, although the degree of fractionation of the metallic parent melt can no longer be accurately inferred. A distinct pallasite parent body is required for the Eagle Station trio, while the origin of the remaining ungrouped pallasites (Glorieta Mountain, Pavlodar) remains uncertain.

2. **Formation:** Core formation may have occurred in two episodes [9]: melting of Fe-FeS eutectic (~1250K), followed by segregation within  $\leq 100\text{K}$  temperature rise; and melting and segregation of metal (~1780K). We assume that mantle silicates and metal remained equilibrated until metal segregated; after that, mantle and core behaved as isolated systems. It is not certain that the extent of partial melting at 1780K (for plausible parent body bulk compositions) is sufficient to exhaust all phases other than olivine. Therefore, regardless of whether the olivine is of cumulate or residual origin, further heating of the parent body may have been required. Further heating would have lowered the fayalite content of the olivine in contact with the core, leading to the conclusion that at the time of metal segregation the mantle may have had a higher in FeO/(MgO+FeO) ratio than 0.12, as observed for main group pallasites [10], but for the following considerations this effect is unimportant. The amount of oxygen dissolved in pure Fe [11] in equilibrium with olivine and pyroxene [12] is given by:

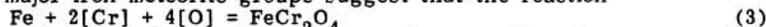
$$\log[\text{O}] = 7.241 - 7629/T + \log(\text{Fa}) \quad (1)$$

where [O] is the oxygen concentration in the melt in ppm. Neglecting the influence of Ni, at  $1780 \pm 30\text{K}$  and  $\text{Fa} = 0.15 \pm 0.05$ , [O] is approximately  $140 \pm 70$  ppm. This is not necessarily the O content of the IIIAB parent melt, because the high-T metallic melt may mix, partially or completely, with the previously formed sulfide melt. Oxygen solubility data for the Fe-FeS eutectic are not available; if they are similar to eqn.(1), its O content, because of its low formation temperature, is negligible. During fractional crystallization of the core [3], O is partitioned into the liquid, since for  $\gamma$ -Fe/liquid [11]

$$\log k_{\text{O}} = -3.31 + 3029/T \quad (2)$$

or  $k < 0.05$  above ~1500K. Thus, like S, O in iron meteorites is derived mostly from trapped liquid, consistent with the occurrence of oxidized minerals in and around troilite nodules [13,16].

Chromium distribution patterns in major iron meteorite groups suggest that the reaction



may limit the O content of the melt: Plots of  $\log \text{Cr}$  vs.  $\log \text{Ni}$  show a steep negative slope, suggesting that

$$[\text{Cr}] = \text{const.} \cdot (1-g)^n \quad (4)$$

with  $n = 1.6 \pm 0.4$  (this value of  $n$  is higher than estimated before [3], but consistent with distribution statistics, assuming that IIIAB irons are an unbiased sample of their parental core). We consider it likely that at liquidus temperatures, O excess controls Cr partitioning: as O gets enriched in the residual liquid Cr is depleted by reaction (3). Taking  $k_{\text{Cr}} = 0.5$  [4], the initial Cr content of the IIIAB melt [14] was  $180 \pm 20$  ppm. This is quite low, considering that the low-temperature melting episode should have extracted >50% of the total parent body Cr [15]. Both lines of evidence suggest that the IIIAB parent melt (and by analogy the parent melts of most major igneous iron meteorite groups) may have been saturated with chromite when it began crystallizing [17], although thermodynamic data are insufficient to assess the feasibility of this model. In any case, the mineralogy and texture of Cape York inclusions indicate that chromite eventually became a liquidus phase [13,16].

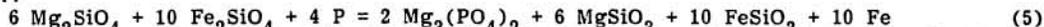
Two other features of Cape York inclusions are noteworthy: (a) not all of the total Cr is present as chromite, but a substantial part is dissolved in troilite, and (b) phosphates are present along with chromite [16]. As T falls in the liquid pocket parental to the inclusion [13], more oxidized phases become stable, some O may diffuse away through the metal (e.g., to form subsolidus precipitates), and the mineral assemblage is eventually controlled by oxygen availability. The environment of trapped melt pockets becomes more oxidizing as fractional crystallization proceeds, so that the most fractionated IIB's contain  $\text{Fe}_3(\text{PO}_4)_2$  [18], whose formation probably requires much higher  $f(\text{O}_2)$  than the mineral assemblage in the Cape York nodules. This may be a combined effect of [O] enrichment, lower crystallization temperatures (because of S and P enrichment [9]), and O diffusion behavior.

3. **Pallasites - Mineralogy:** Pallasite formation differs from the evolution of liquid pockets in irons in that O is buffered by O-bearing minerals. Pyroxene, although rare, seems to be present in all main group pallasites [10], and hence eqn.(1) is probably valid. The metallic liquid from which main group pallasites formed had undergone extensive fractional crystallization, and therefore, compared to the time of segregation (a) its content of S and P was much higher, and (b) T much lower. Under these conditions, phosphate formation is favored [19], accompanied by reduction of FeO. Pallasites contain phosphates of the type  $\text{M}_3(\text{PO}_4)_2$ : farringtonite ( $\text{M}=\text{Mg}$ ), stanfieldite ( $\text{M}=\text{Ca}_4(\text{Mg},\text{Fe})_5$ ), and merrillite ( $\text{M}=\text{Ca}$ ;

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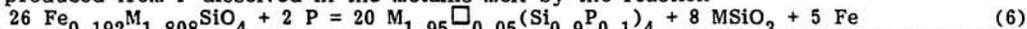
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formerly "whitlockite") [21]. Phosphorus was presumably provided by the metallic melt, cations by the silicates, e.g.:



Sparingly chalcophilic elements dissolved in the metallic melt, which form phases of the type  $\text{NaMPO}_4$  in Cape York nodules [16], are present as traces in pallasitic phosphates. Since all phosphate-forming reactions also form pyroxene, it is difficult to tell whether some pyroxene is primary (i.e., was present before core liquid and mantle were brought in contact). Its low Ca [10] is not unequivocal proof of secondary origin, since  $\text{Ca}_3(\text{PO}_4)_2$  is very stable [18], and reaction with a metallic melt high in P will significantly deplete Ca from the silicates. Textures of pyroxene occurrence [10] do indicate a secondary origin.

The members of the Springwater quartet appear to have been formed from a somewhat more evolved, but generally similar metallic liquid [1,2]; Fa is more variable, but higher than in main group pallasites. Three of them (and one member of the main group) contain phosphoran olivine [10,20]; their phosphate mineralogy seems slightly more diverse, including the apparent disequilibrium assemblage farr-stan-merr not found in the main group [21]. Phosphorus (and probably S) in the metallic melt may have been higher, and bringing it in contact with somewhat more Fe-rich olivine results in an assemblage even more removed from equilibrium than in the case of main group pallasites. The formation of phosphoran olivine can be understood, analogous to phosphate formation, as redox reaction involving reduction of FeO from olivine. If P replaces Si, and charge balance is maintained by cation vacancies [20], phosphoran olivine is  $\text{M}_{2-x/2}\text{O}_{1-x/2}(\text{Si}_{1-x}\text{P}_x)\text{O}_4$ , where  $\text{M}=(\text{Mg},\text{Fe})$ , and  $x \sim 0.1$  [10]. Phosphoran olivine can be produced from P dissolved in the metallic melt by the reaction



The phosphoran olivine produced would have a significantly lower Fe/Mg ratio than the P-free parent olivine, but rapid Fe Mg exchange can be expected at the temperatures involved, particularly if our supposition of constitutional cation vacancies is correct.

4. Pallasites - Origin: If mantle and core were in continuous contact, an equilibrium assemblage should result. If, on the other hand, melt was injected into an olivine layer catastrophically, the system may initially have been very far from equilibrium, and could conceivably have reached a number of metastable assemblages depending on phase nucleation, local compositional fluctuations, etc. Unless the farr-stan-merr assemblage in the Springwater quartet is the result of stabilization by minor elements [21], equilibrium was not maintained on the scale of a few cm, arguing for a catastrophic origin [2]. The situation is less clear for main group pallasites, which do not contain disequilibrium phosphate assemblages. The general similarity between main group pallasites and the Springwater quartet (except for the S-rich member, Phillips County) makes different modes of origin implausible. Moreover, even in the absence of thermodynamic data it seems quite unlikely that phosphoran olivine, which occurs in at least one main group member in addition to Springwater, Rawlinna, and Zaisho, is a stable phase. We therefore favor an origin of main group pallasites and the Springwater quartet by two separate injection events, most likely in the same parent body. The difference in Fa implies either a laterally inhomogeneous mantle, or injection into a higher layer (i.e., a cumulate from a more evolved liquid) in the case of Springwater.

Metal compositions, especially Ni contents, scatter considerably among pallasites. We consider this primarily the result of nucleation effects in the liquid after injection into the olivine layer, although this does not readily explain the Ni-Fa correlation [1]. Evidently, cooling following injection was controlled by variations in the immediate environment, causing variations in olivine morphology between individual pallasites [22], whereas the metallography [23] apparently reflects the asteroidal cooling rate.

The behavior of P is crucial to the argument of Davis [8], who pointed out that phosphides are less abundant in pallasites than they ought to be if pallasites originated from a liquid in equilibrium with fractionated IIIB irons, and a solid/liquid partition coefficient  $k_p=0.12$ . Considering the above scenario, this does not necessarily argue against an origin from the IIIAB core liquid. First, phosphates may have been molten when they formed [10,21], and some immiscible phosphate melt may have been lost. Second, in the evolved metal liquids from which main group and Springwater pallasites formed, S was probably high [9], and  $k_p \sim 0.5$  rather than 0.12 [7]. The sulfide abundance in Phillips County clearly demonstrates that it formed from a S-rich melt. The fact that we do not have more S-rich pallasites is most probably due to sampling bias [9].

In summary, it is very likely that main group pallasites and IIIAB irons come from the same parent body. More detailed study is needed to decide whether the Springwater quartet also formed on this body.

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