

In the infrared, the spectra of CM chondrites Murray and Murchison are quite similar, with broad absorptions at 3  $\mu\text{m}$  and from 8 to 12  $\mu\text{m}$ , with a narrower feature centered on 6.2  $\mu\text{m}$ . The Mg serpentine, amesite, has abundant spectral features beyond 13  $\mu\text{m}$  that are not seen in the CM chondrite spectra. The Fe serpentines have absorptions that can contribute to those seen in the CM chondrites, but lacks the large absorptions beyond 13  $\mu\text{m}$ , again providing a better spectral match than the Mg serpentines.

In the future we hope to compare the spectra of these Fe serpentines with a wider variety of CM chondrites. Additionally, a theoretical modeling study is planned that will attempt to match meteorite spectra using their mineralogy and grain size distribution as the initial input to the models.

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**References:** [1] Zolensky M. and McSween H. Y. (1988) *Meteorites and the Early Solar System* (Kerridge and Matthews, eds.), 114–143. [2] King T. V. V. and Clark R. N. (1989) *JGR*, 94, 13997–14008. [3] Fleischer M. and Mandarino J. A. (1991) *Glossary of Mineral Species*. [4] Vilas F. and Gaffey M. J. (1989) *Science*, 246, 790–792.

**THERMAL HISTORIES OF PGE-RICH METAL PARTICLES IN A VIGARANO CAI.** I. Casanova<sup>1,2</sup> and L. Grossman<sup>2,3</sup>, <sup>1</sup>Department of Geology, Field Museum of Natural History, Chicago IL 60605-2496, USA, <sup>2</sup>Department of the Geophysical Sciences, University of Chicago, Chicago IL 60637, USA, <sup>3</sup>Enrico Fermi Institute, The University of Chicago, Chicago IL 60637, USA.

Metal particles in Vigarano 1623-8, a type B2 CAI [1], underwent virtually no sulfidation, as is typical of opaque assemblages from Ca-Al-rich inclusions in the reduced CV3 chondrites [2]. In this study, we have identified two large metal grains (M1 and M2) with chemical and mineralogical features that may indicate cooling under different conditions and are, therefore, difficult to understand in the environment of a single CAI's thermal evolution. M1 is an almost spherical, kamacite+taenite-bearing particle included in a fassaite grain of the CAI host; a 17.5- $\mu\text{m}$ -long (0.5- $\mu\text{m}$  steps) microprobe traverse along M1 shows that Ni and Ru contents in the taenite (31.5 wt% and 1.1 wt% respectively) are uniform, and differ from those in the adjacent kamacite (Ni = 4.5 wt%, Ru = 0.7 wt%).

M2 is a 20- $\mu\text{m}$ , irregularly shaped taenite particle embedded in a fine-grained (spinel-rich) portion of 1623-8. It has a homogeneous composition with 10.5 wt% Ni, 0.4% Co, 0.7% Re, 0.6% Pt, and high concentrations of Ru (6.5 wt%), Os (4.3 wt%), and Ir (8.2 wt%), as previously recognized by [1]. The composition of M2 is such that it should have undergone exsolution at  $800 \geq T \geq 600^\circ\text{C}$  (according to experimental data by [3]) to form at least two ( $\alpha$ + $\gamma$ -NiFe), or probably three (+ $\epsilon$ -RuFe) different phases. Lack of exsolution features in this large grain is therefore indicative of equilibration at relatively high temperatures ( $T > 600^\circ\text{C}$ ) followed by rapid cooling. Other metal particles of similar bulk compositions in CAIs from the Leoville chondrite (also a reduced CV3) show extensive exsolution features that have been interpreted as the result of low-temperature equilibration of the CAIs and their constituents after incorporation into their parent body [4,5].

The relatively high equilibration temperature of M2 is, however, inconsistent with the existence of kamacite in M1. From the phase relations in the Fe-Ni binary, a grain like M1, with 25 wt% bulk Ni, would barely start to exsolve kamacite at  $550 \pm 22^\circ\text{C}$  (for a  $25 \pm 10$  vol% kamacite fraction, assuming that the effect of 2.5 wt% Co and  $\sim 3$  wt% total PGE is not substantial). Therefore, exsolution of kamacite in M1 must have occurred prior to its incorporation into 1623-8. Then, in order to preserve the sharp taenite-kamacite boundary observed in M1, the cooling rate of the inclusion at  $T \leq 1200^\circ\text{C}$  must have been fast since diffusion of FeNi at  $50^\circ\text{C/hr}$  in the  $1200^\circ\text{C}$ – $550^\circ\text{C}$  range would result in appreciable reequilibration of kamacite and taenite after just  $\sim 1$  hr. This is in conflict with previous experimental results that suggest that the reverse zoning observed in melilites from type B CAIs formed at cooling rates  $\leq 50^\circ\text{C/hr}$  [7]. Then, M1 could not have been incorporated into the CAI during the melting event that produced the primary melilites in 1623-8 (with reverse zoning [1]), but could have entered the inclusion during a second melting stage, such as proposed by [1], followed by very fast cooling ( $> 50^\circ\text{C/hr}$ ).

After kamacite exsolution, the cooling history of M1 at  $T \leq 550^\circ\text{C}$  must have been slow. Calculations based on  $D_{\text{Ni}}^{\text{Taenite}} \sim 10^{-18} \text{ cm}^2 \text{ s}^{-1}$  [6] show that the homogenization of Ni in taenite over distances of 5  $\mu\text{m}$  (as observed in M1) would take about 8000 yr at  $550^\circ\text{C}$  (or require a cooling rate  $\sim 0.1^\circ\text{C/yr}$ ). This should be considered an upper limit to the cooling timescale since the probable effect of the few percent levels of Co and PGEs in this grain will be to accelerate Fe-Ni interdiffusion, and taenite may homogenize substantially faster than in the pure binary system. But, if such slow cooling had taken place in the CAI, M2 (just a few hundreds of micrometers apart from M1) would have undergone exsolution [3]. This means that not only kamacite exsolution but also production of homogeneous taenite in M1 predate its incorporation into the CAI. The phase assemblages in M1 and M2 cannot be reconciled with a single thermal history and thus require separate environments for their early thermal evolution.

**References:** [1] MacPherson G. J. and Davis A. M. (1993) *GCA*, 57, 31–243. [2] Casanova I. and Grossman L. (1993) *LPS XXIV*, 257–258. [3] Blum J. D. et al. (1989) *GCA*, 53, 483–489. [4] Simon S. B. and Grossman L. (1992) *EPSL*, 110, 67–75. [5] Blum J. D. et al. (1989) *GCA*, 53, 543–556. [6] Dean D. C. and Goldstein J. I. (1986) *Met. Trans.*, 17A, 1131–1138. [7] MacPherson G. J. et al. (1984) *J. Geol.*, 92, 289–305.

**PRIMORDIAL MINERAL GROWTH IN A PLASMA.** W. A. Cassidy and C. M. Kern, University of Pittsburgh, Pittsburgh PA 15260, USA.

Diamonds, SiC, TiC, corundum, and graphite have been found in primitive meteorites [1] as crystalline components existing with, but far removed from, equilibrium with lower-temperature minerals and, in some cases, hydrocarbons. From anomalous isotopic ratios there is a presumption that the grains, many of them submicroscopic in size, had an extrasolar system origin [1] and may have formed within stellar atmospheres. If that were true, their environment of formation was a plasma. A plasma is an electrically neutral gas containing a high proportion of its component molecules in the ionized state, with charge balance satisfied by the presence of free electrons; it has been described as a fourth state of matter. The exact