Met. Planet. Sci. 31 (1996) Suppl.

and CV3 matrixes and conclude instead that these minerals formed by fluid-rock interaction during asteroidal alteration.

Salite-hedenbergite-andradite-wollastonite-kirschsteinite assemblages are abundant in Allende DIs, which appear to have experienced various degrees of aqueous/hydrothermal alteration and subsequent dehydration during thermal metamorphism [6–8]. These assemblages form concentrically zoned, irregularly shaped inclusions and veins in matrixes of the DIs and complex rims around them. Pyroxene-andradite inclusions (10–50 µm in diameter) are texturally and compositionally similar to those in matrixes of the oxidized CV3s, suggesting a common origin. Both occurrences consist of pyroxene layers and andradite cores; Ni-rich metal and sulfides are minor. Pyroxenes are porous and highly variable in composition ($Fs_{10}Wo_{50}$–$Fs_{40}Wo_{50}$) on a submicrometer scale. The outer pyroxene layers are more magnesian than the inner layers, and andradite is pure $Ca_3Fe_2Si_3O_{12}$. X-ray imaging of the DIs and oxidized CV3s reveals that their matrixes are uniformly enriched in Ca, which is mainly concentrated in pyroxene-andradite inclusions, whereas chondrules and chondrule pseudomorphs are significantly depleted in Ca. This suggests that Ca was lost from the chondrules during alteration and precipitated in pyroxene-andradite inclusions in matrixes of the DIs and oxidized CV3s. Supporting evidence comes from the reduced CV3s, where matrixes lack pyroxene-andradite inclusions and chondrules are unaltered and contain anorthitic mesostases.

Pyroxene-andradite veins (up to 3 mm long) in DIs have similar mineralogy, zoning (from periphery to center: salite → ferrosalite → andradite), and compositions of pyroxenes and andradite to those in matrix inclusions, suggesting a common origin. Veins are commonly observed near edges of the DIs where they connect with the pyroxene-andradite rims. Veins replace fayalitic rims around isolated forsterite grains, indicating vein formation after formation of the fayalitic rims. Rare pyroxene-andradite veins occur in the Allende host as well.

The best-developed pyroxene-andradite rims (~100 µm wide) occur around heavily altered DIs, i.e., with chondrules that are completely replaced by fayalitic olivine. The rim layers from inside to outside include salite → hedenbergite → andradite + wollastonite + kirschsteinite → ferrosalite. Wollastonite and salite-ferrosalite pyroxenes may form needles and prismatic crystals. The outer portions (~0.5–1 mm) of the rimmed DIs are significantly depleted in Ca, suggesting that Ca was leached from the DIs and precipitated in veins and rims at the boundaries with the host Allende. The Allende matrix around the DIs contains intrusions of rim material and is also depleted in Ca, indicating that the rims were partly or entirely formed *in situ*. The coexistence of andradite, wollastonite, and kirschsteinite in the rims constrains the lower limit of the O fugacity during rim formation [$2Ca_3Fe_2(SiO_4)_3 = 2CaSiO_3 + 4CaFeSiO_4 + O_2$], which is several orders of magnitude higher (between 300 and 800 K) than the solar nebular gas.

We conclude that these Ca-Fe-Mg silicate assemblages in Allende DIs, Allende host, and other oxidized CV3s formed by fluid-assisted metamorphism in an asteroidal environment. The reduced CV3s largely escaped this alteration. We infer that (1) the evidence for a highly oxidized solar nebula is weak, (2) matrixes in oxidized CV3s are not primitive, and (3) absolute dating of secondary phases in CAIs dates asteroidal alteration.

References: [1] Peck J. (1985) Bachelor's dissertation, Harvard Univ., 123 pp. [2] MacPherson G. et al. (1988) in *Meteorites and the Early Solar System* (J. Kerridge and M. Matthews, eds.), pp. 746–807, Univ. of Arizona, Tucson. [3] Ikeda Y. and Kimura M. (1995) *Proc. NIPR Symp. Antarct. Meteorites, 8*, 819–831. [4] Kojima T. and Tomeoka K. (1994) *LPS XIII*, 399–400. [5] Krot A. et al. (1995) *Meteoritics, 30*, 748–775. [6] Krot A. et al. (1996) *LPS XXVII*, 711–712.

DETERMINATION OF POROSITY OF METEORITES—COMPARISON OF TWO METHODS.

K. Kuoppamäki[1], L. Kivekäs[2], J. Timonen[1], J. Hartikainen[1], K. Hartikainen[1], and L. J. Pesonen[2], [1]Department of Physics, University of Jyväskylä, FIN-40351 Jyväskylä, Finland, [2]Department of Geophysics, Geological Survey of Finland, FIN-02150 Espoo, Finland.

Porosity is an important physical property in meteorite research and in studies of meteorite impact craters. For example, the calculation of the Hugoniat curves for shock recovery experiments require knowledge of the porosity of the sample [1]. Physical properties of meteorites have been used to rapidly classify meteorites into main classes and groups [2]. In this method porosity causes scatter in the bivariate diagrams of the physical properties and thus reduces the application of the method. In distinguishing the primary and secondary consolidation states of the meteorite parent bodies, the nature of porosity of meteorites is of vital importance [3]. On the other hand, the construction of geophysical (e.g., gravity, seismic) models of impact craters requires knowledge of the porosity of the rocks [4,5]. Porosity is relatively easy to measure by the classical water immersion technique [6], which is based on the mass difference between water-saturated and dried samples. However, many museums do not allow water to be used in measuring meteorite samples, and for this reason we have developed a new He-pycnometer device to measure accurately the porosity of meteorites without changing their physical or chemical characteristics. In this paper we present a comparison of He-pycnometer porosity data with those obtained by the water immersion technique for 20 meteorite samples.

The He-pycnometer device consists of two chambers, the reference chamber (volume Vr) and the sample chamber (volume Vs), and a number of pressure (P) and temperature (T) gauges. First, the pressure and temperature of the reference chamber filled with He (Pr, Tr) are measured, after which the He is allowed to expand to the sample chamber, and the resulting Ps and Ts are recorded and time-monitored to obtain stationary readings. Knowing the geometric volume of the sample and the residual pressure of the chambers before they are filled with He, the porosity (ε) of the sample can be calculated using ideal gas equations [7]. The system is calibrated with standard Al disks. The advantages of this method are that measurements are relatively fast, and the often rare and fragile meteorite samples are not disturbed (weathered) by water.

In order to check the reliability of the instrument we compared He-pycnometer porosities with those obtained by the classical water immersion technique [6,7]. In this comparison we have used various geological rock samples (in addition to meteorites) with variable porosity, mass, and shape. The water immersion measurements were carried out in the GSF laboratory by soaking the samples in tap water for 2–5 days and then drying the samples in an oven (≤105°C) for a few days. The pore volume is calculated from the mass difference between water-saturated and dried samples, and the sample volume is determined by weighing the surface-dried wet sample in air and water (Archimedes' principle).

Comparison of the two datasets reveals that there is an overall agreement with porosity over a wide range of values from 0.5% to 30%. However, the porosity values do not plot exactly on the 1:1 relation line: The He porosities appear systematically lower than those obtained with the water-soaking technique. We will discuss the new porosity data and their error sources and present some applications of the data for meteorite research.

References: [1] Schmitt et al. (1994) *LPS XXV*, 1209–1210. [2] Pesonen L. J. (1994) *Proc. NIPR, 6*, 401–416. [3] Yomogida K. and Matsui T. (1983) *JGR, 88*, 9513–9533. [4] Kukkonen I. et al. (1992) *Tectonophysics, 216*, 111–122. [5] Plado J. et al. (1996) *Meteoritics, 31*, 289–298. [6] Kivekäs L. (1993) *GSF Spec. Paper 18*, 119–127. [7] Kuoppamäki K. (1996) M.Sc. thesis, in preparation, Univ. of Jyväskylä, Finland.

A CHONDRULE MICROMETEORITE FROM ANTARCTICA WITH VAPOR-FRACTIONATED TRACE-ELEMENT ABUNDANCES.

G. Kurat[1], P. Hoppe[2], and C. Engrand[1], [1]Naturhistorisches Museum, A-1014 Vienna, Austria, [2]Physikalisches Institut, Universität Bern, CH-3012 Bern, Switzerland.

Chondrules are rare among micrometeorites (MMs) but apparently do contribute to the cosmic spherule (CS) population [e.g., 1]. This probably is caused by the relatively large masses of chondrules that prevent atmospheric entry without fusion. However, a few unmelted MMs do appear to be original chondrules. One such object was found in mount 94-4 (#32). This chondrule was studied in detail by utilizing optical microscopy, scanning electron microscopy, electron microprobe, and ion microprobe techniques.

Chondrule 94-4-32 is a ~80% fragment of an originally oval-radiating olivine chondrule about 120 µm in diameter. Olivine has ~Fa_{35}, is platy, and is only a few micrometers thick, but up to about 50 µm long. Between the plates are variable amounts of Na,Ca,Al-rich glass and some open pore space, a few clinopyroxenes, and submicrometer grains of pentlandite. The

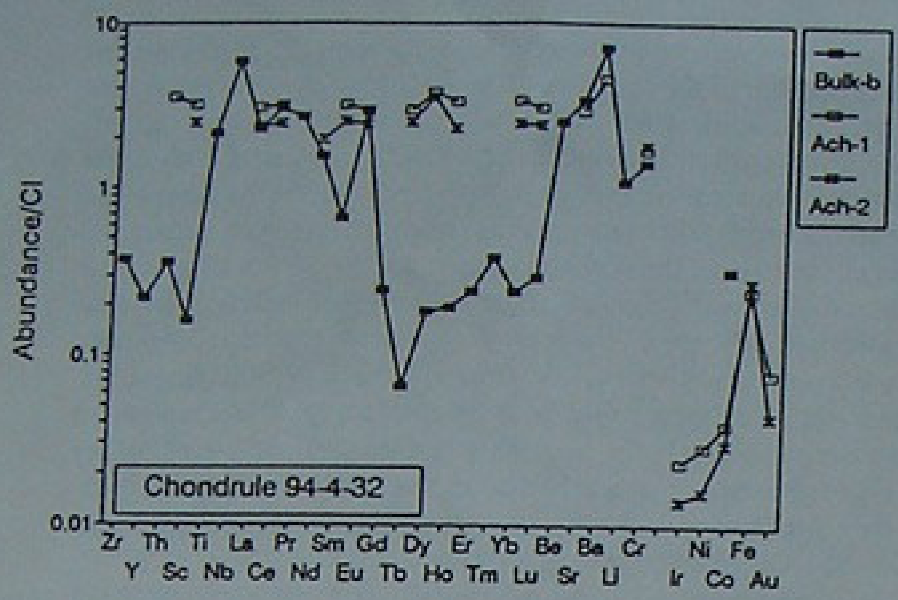

Fig. 1. Normalized [7] trace-element abundances in chondrule 94-4-32 (as measured by SIMS) compared to those of two RP chondrules from Allende, Ach-1 and Ach-2 [5].

bulk composition of 94-4-32 is principally chondritic but enriched in CaO (3.8 wt%) and depleted in NiO (0.02 wt%), K_2O (<0.02 wt%), and S (SO_3 = 0.33 wt%) compared to chondrites. However, trace-element abundances (Fig. 1) are distinctly different from those in chondrites. All highly refractory elements (Zr, Y, Th, Sc, Sm, Gd, Tb, Dy, Ho, Er, Tm, Yb, and Lu) have subchondritic abundances (0.007–0.7× CI), whereas the moderately refractory elements (Ti, Nb, La, Ce, Pr, Nd, Eu, Be, Si, and Ba) have superchondritic abundances (~2–7× CI). The moderately volatile elements Li and Cr have approximately chondritic abundances, and Co and Ni have subchondritic abundances.

The depletion of 94-4-32 in siderophile elements and the presence of pentlandite can be taken as additional indications for that particle being a chondrule. Fractionated lithophile-element abundances have been observed in chondrules from a few chondrites [e.g., 2,3] and are probably due to sampling fractionation. Chondrule 94-4-32, however, has lithophile-element abundances that clearly reflect vapor fractionation, comparable to that found in type II CAIs [4], and thus gives evidence for the formation of that chondrule by condensation from a vapor that was depleted in superrefractory elements. A condensation origin of some chondrules, like the RP chondrules, has been shown to be likely because of vapor-fractionated siderophile-element abundances [5]. However, chondrule 94-4-32 is the first chondrule found so far that carries an unequivocal signal of condensation in its lithophile-element abundance. Chondrule 94-4-32 and CSs with similar trace-element patterns [1] indicate that such matter is apparently more abundant in the interplanetary dust, which bears compositional similarities to CM/CR chondrites [6], than in chondrites, but what does this imply?

Acknowledgments: Support by the FWF in Austria and the Schweizerische Nationalfonds in Switzerland and help with EMPA by J. Walter is gratefully acknowledged.

References: [1] Walter J. et al. (1995) *LPS XXVI*, 1457. [2] Kurat G. (1984) *Proc. 27th IGC, 11*, 155. [3] Kurat G. et al. (1983) *EPSL, 68*, 43. [4] Martin P. M. and Mason B. (1974) *Nature, 149*, 333. [5] Kurat G. et al. (1985) *LPS XVI*, 471. [6] Kurat G. et al. (1994) *GCA, 58*, 3879. [7] Palme H. et al. (1981) in *Landolt-Boernstein* (K. Schaifers and H. H. Voigt, eds.), p. 257, Springer-Verlag.

SHOULD WE EXPECT TO FIND MORE PIECES OF THE K/T BOLIDE?

F. T. Kyte, Institute of Geophysics and Planetary Physics, University of California, Los Angeles CA 90095-1567, USA.

The discovery of a small (~2.5 mm), heavily altered, meteorite fragment [1] in K/T boundary sediments (DSDP 576, North Pacific) raises many questions. These include (1) Can this be a piece of the K/T bolide ejected from the Chicxulub impact? and (2) Do other K/T meteorites exist that can be recovered?

The answer to the first question is clearly yes, based on theoretical, experimental, and observational studies. Theoretical models of impact cratering [e.g., 2] describe waves of rarefaction on the backside of impacting projectiles that can allow spallation of projectile material that does not experience peak shock pressures. Early results from numerical simulation of the Chicxulub impact [3] indicate that several percent of the projectile may have survived melting. Experimental studies of low-angle (<30°) impacts [4] show that large portions of a projectile can ricochet downrange, greatly enhancing the potential for survival of unmelted meteorites. Actual examples of meteorite survival following hypervelocity impact are rare. Although iron meteorites have been recovered near several small craters, they may have survived by separation prior to the impact. The only well-documented case of meteorite survival during hypervelocity impact is in the debris field of the Late Pliocene impact of the Eltanin asteroid into the 5-km-deep ocean in the sub-Antarctic Pacific. Recoverable impact debris (mostly impact melt) is entirely meteoritic material because the target was seawater [5]. Several percent of this debris is unmelted meteorite fragments, commonly lithic clasts 1–4 mm in size, as well as unmelted grains and lithics included within the impact melt. This debris extends across at least 600 km of the ocean floor [6] and the mass of unmelted meteorites could easily exceed 10^{12} g.

The answer to the second question is maybe and depends on whether the K/T meteorite is, in fact, directly related to the K/T impact. Based on a single sample, one cannot rule out the unlikely possibility this is just a stray micrometeorite. The best proof that it is from the impact would be to find another piece at another site, thus providing a more affirmative answer to the question. The apparent absence of K/T meteorites, despite 15 years of detailed research, may reflect a bias in sampling methods and poor preservation at most K/T localities. The DSDP 576 meteorite was nearly entirely altered to clays and hydrated Fe-oxides. At most K/T localities where compaction and lithification of thick sedimentary sequences is typical, such an object may be crushed and indistinguishable from typical clayey matrix. Further, sample-processing procedures for microfossil, mineralogical, or geochemical analyses usually involve sample disaggregation, crushing, or chemical leaching that would destroy these meteorites. The DSDP 576 meteorite was recovered because this thin sedimentary sequence has barely been compacted since its deposition and the meteorite happened to be at the cut surface of the core, where it was identified as an anomalous particle within the surrounding clay-sized sediment.

Clearly the best opportunities for finding other K/T meteorites would be in other unconsolidated deep-sea sediments. How many should we expect? The DSDP 576 meteorite contains only ~0.1% of the Ir in that core. If only 1% of the K/T bolide were ejected as similar-sized debris, we might expect 10 pieces in this core. Actually, three cores were taken at site 576, three at 577, and several others exist in the Pacific. The next question is, how do we get them?

References: [1] Kyte F. T. (1996) *LPS XXVII*, 717. [2] Melosh H. J. (1989) *Impact Cratering*, Oxford Univ., 245 pp. [3] Pierazzo E. et al. (1996) *LPS XXVII*, 1029. [4] Schultz P. H. and Gault D. E. (1990) *GSA Spec. Paper 247*, 239. [5] Kyte F. T. and Brownlee D. E. (1985) *GCA, 49*, 1095. [6] Kyte F. T. et al. (1988) *Science, 241*, 63.

STUDY OF IMPACT Ni-RICH MAGNESIOFERRITE SPINELS UNDER TEM.

F. Langenhorst and P. Claeys, Institut für Mineralogie, Museum für Naturkunde, Invalidenstrasse 43, D-10115 Berlin, Germany.

Millimeter-sized spherules, now altered to clay minerals, are reported in many Cretaceous-Tertiary (K/T) boundary sites [1,2]. Two types of spherules can be distinguished based on their morphology, the presence of high-temperature phases, their geographic distribution, and their stratigraphic position within the K/T layer [3,4]. Rounded (but also elongated or teardrop-shaped) spherules, some with a preserved glass core reflecting target rock composition [5–7], are found only at proximal sites at the base of the K/T layer. The second type, similar to microkrystites, is found worldwide, associated with Ir in the uppermost part of the K/T layer. These spherules are always spherical and typically contain a high-temperature phase in the form of Ni-rich magnesioferrite spinels [8–10]. This study deals only with this second type of spherule, in particular the magnesioferrite phase. Similar magnesioferrite-bearing spherules are found in meteorite fusion crust, cosmic spherules [10], sediments from the Eocene-Oligocene boundary in Massignano (Italy), and Late Pliocene sediments containing debris from a