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ison et al. (1987) *GCA, 51*, 1875–1882. [3] Fredriksson et al. (1989) *Zeit. Natur., 44a*, 945–962. [4] Christoffersen R. and Buseck P. P. (1983) *Science, 223*, 56–58. [5] Bradley et al. (1984) *Science, 223*, 1327–1329. [6] Sugiura et al. (1984) *Proc. LPSC 14th*, in *JGR, 91*, B641–B644. [7] Mostefaoni S. and Perron C. (1994) *LPS XXV*, 945–946.

COARSE-GRAINED RIMS ON MAGNESIUM-RICH AND MAGNESIUM-POOR CHONDRULES IN ORDINARY CHONDRITES.

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Chondrules with igneous rims and enveloping compound chondrules [1] in ordinary chondrites (OC) preserve the record of solids that were present at different times or in different regions of the solar nebula during chondrule formation. These objects demonstrate that OC chondrules experienced multiple episodes of chondrule formation. This conclusion is consistent with (1) the presence of relict grains in chondrules [2,3] that were probably produced by disaggregation of chondrules of a previous generation, (2) small range in OC chondrule O isotope composition [4] (which may have resulted from homogenization during several chondrule melting episodes), and (3) interelement correlations in bulk chondrule data that can be interpreted as the random sampling of a previous generation of chondrules [5].

Rims around chondrules can be divided into two major categories: fine-grained rims (FGR), typically opaque and Fe rich [6,7], and relatively coarse-grained rims (CGR) [8]. FGR are similar in mineralogy, chemistry, and grain size to matrix material and consist of micrometer- to submicrometer-sized grains or amorphous material. It is widely accepted that the FGR accreted around chondrules in the solar nebula and were not subsequently heated before aggregation into a parent body [7]. CGR are coarser than matrix material and surround ~10% of the chondrules in OC [8]. Rubin [8] found that CGR in OC consist mainly of olivine and suggested that these rims formed by sintering of opaque matrix material by heating to subsolidus or subliquidus temperatures during chondrule formation.

Thirteen CGR on Mg-rich chondrules (type I, Fa/Fs < 10 mol%) and nine rims on Mg-poor chondrules (type II, Fa/Fs > 10 mol%) were studied petrographically, by electron microprobe analysis, and scanning electron microscopy. Many of the CGR on type I chondrules show evidence of significant and, in many cases, complete melting including (1) resorbed chondrule surfaces along the boundaries with CGR, (2) presence of rounded metal/troilite nodules and feldspathic mesostasis, (3) euhedral and subhedral morphology of olivine and pyroxene, and small variations in their grain sizes, and (4) absence of relict grains. These rims can be classified as igneous rims. Although the majority of type I host chondrules studied have olivine as a major phase, pyroxene is a predominant phase in many of their igneous rims. Olivine and pyroxene in the rims (Fa_{1-6}, Fs_{2-9}) are compositionally similar to those of the host chondrules ($Fa_{0.5-7}$, Fs_{1-7}). CGR on type I chondrules are typically surrounded by accretionary FGR; the boundary between these rims is abrupt. Type II host chondrules and their CGR have similar modal mineralogy and mineral chemistry (Fa_{10-34}, Fs_{10-30}, and Fa_{16-44}, Fs_{10-31} respectively). Although these rims also show evidence of significant heating, in most cases these rims were not completely melted; many contain unmelted Mg-rich relict grains or chondrule fragments (Fa_{0-10}, Fs_{1-9}).

Similar Fa and Fs contents in mafic minerals of OC igneous rims and their type I chondrule hosts indicate that many OC chondrules experienced multiple heating events during a time short compared to the time necessary for appreciable evolution in the mean Fa or Fs of the nebular solids, and were then withdrawn from the chondrule-forming region. Type II chondrules and their CGR formed from more oxidized material mixed with fragments of type I chondrules and were heated to lower temperatures than type I chondrules and their CGR. Type I and type II chondrules may have formed in different OC nebular subregions or at different times and were mixed together before or during agglomeration to form chondrites.

References: [1] Wasson et al. (1994) *GCA*, submitted. [2] Rambaldi E. R. (1981) *Nature, 293*, 558–561. [3] Nagahara H. (1981) *Nature, 292*, 135–136. [4] Clayton et al. (1991) *GCA, 55*, 2317–2339. [5] Alexander C. M. O'D. (1994) *LPS XXV*, 7–8. [6] Scott et al. (1984) *GCA, 48*, 1741–1757. [7] Alexander et al. (1987) *Meteoritics, 22*, 316–317. [8] Rubin A. E. (1984) *GCA, 48*, 1779–1789.

^{40}Ar-^{39}Ar AGES OF DIFFERENT LITHOLOGIES FROM THE UNIQUE EUCRITE PADVARNINKAI.

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Padvarninkai was recently described as a polymict eucrite containing three different lithologies: Pad-1, fine- to coarse-grained eucritic clasts; Pad-2, a fine-grained lithology with quenched texture; and Pad-3, a partly glassy matrix between eucritic clasts [1]. In addition to Pb-Pb studies on zircons from lithology Pad-1—crystallization age 4.55 b.y. [1]—we analyzed all three lithologies by high-resolution stepwise heating ^{40}Ar-^{39}Ar dating to determine the postcrystallization thermal history of the meteorite.

Pad-1 yields a well-defined plateau age of 3.80 b.y. for the high-temperature fractions (1190°–1530°C) comprising 65% of ^{39}Ar released. This age marks the time when the rock was totally degassed by, most likely, impact-induced reheating. In the first 35% of the spectrum the apparent ages are lower indicating partial loss of $^{40}Ar_{rad}$. However, the very first fractions (450°–750°C) point toward the presence of $^{40}Ar_{exc}$ since the apparent ages of 2.3–3.3 b.y. clearly exceed the value for the youngest fraction, 0.6 b.y. at 950°C. The lowest value provides an upper limit to the last time when the rock was subjected to thermal strain responsible for partial degassing. Two explanations are possible for the occurrence of $^{40}Ar_{exc}$ in low-temperature fractions: (1) Shock causes redistribution of $^{40}Ar_{rad}$ as demonstrated by analyses of experimentally shocked rocks [2,3]. Another hint toward shock metamorphism comes from the K/Ca spectrum: While unshocked eucrites usually yield nearly constant K/Ca ratios in the first ~70% of ^{39}Ar released from degassing homogeneous plagioclase [4], Pad-1 has monotonously decreasing K/Ca ratios for the first 30% ^{39}Ar released, very similar to experimentally shocked gabbro [2]. (2) There is evidence for atmospheric contamination of the low-temperature fractions from an isochron with $^{40}Ar/^{36}Ar \sim 295.5$ (air). With this correction the apparent ages shift down close to 0.6 b.y.

Pad-3 yields a very similar degassing pattern, apparent age, and K/Ca spectra to Pad-1, but no well-defined high-temperature plateau. The apparent ages for the 1180°–1530°C fractions, which again contain ~65% ^{39}Ar, increase monotonously from 3.44 to 3.71 b.y. Thus, the glassy matrix of Pad-3 was most likely formed 0.6 Ga by melting of Pad-1-like rock where also the unmolten eucritic lithology Pad-1 was affected. Subsequently both lithologies cooled rapidly as their K-Ar systems are rather weakly disturbed.

Pad-2 is less retentive than the two others: At 1000°C already 65% of ^{39}Ar is released compared to 19% and 14% for Pad-1 and Pad-3 respectively. After correction for a trapped component the first 65% of the age spectrum has a plateau at 1.71 b.y. The isochron for the fractions with trapped ^{36}Ar yields a slope with the same age. For the last 35% of the spectrum the ages increase up to ~3 b.y. Pad-2 contains maskelynite of bytownitic composition reflecting severe shock and a quenched texture indicating rapid cooling to subsolidus temperatures. The plateau age most likely dates these events. As even severe shock itself cannot completely reset the K-Ar system [5] we conclude that the subsolidus cooling was slow enough to cause major $^{40}Ar_{rad}$ loss. After 1.71 b.y. the K-Ar system remained completely closed. Therefore, it is impossible that lithology Pad-2 was in thermal contact with Pad-1 and Pad-3 since these lithologies were partially degassed as late as 0.6 Ga. Thus, 0.6 b.y. is an upper limit for the compaction age of the Padvarninkai eucrite.

References: [1] Bukovanská M. et al. (1991) *Meteoritics, 26*, 325. [2] Trieloff M. et al., this volume. [3] Bogard D. et al. (1987) *GCA, 51*, 2035–2044. [4] Kunz J. et al. (1994) *Space Planet. Sci.*, special issue, submitted. [5] Stephan Th. and Jessberger E. K. (1992) *GCA, 56*, 1591–1605.

OXYGEN ISOTOPES IN SPINELS FROM ANTARCTIC MICROMETEORITES.

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Spinel-rich inclusions were found in a large unmelted micrometeorite (MM) from Antarctica [1]. This particle (MM92/15-23) consists of a fine-grained matrix of dehydrated former phyllosilicates that enclose a few small olivines, one large chromite, and several spinel-rich inclusions. The latter form elongated to rounded bodies up to 35 µm in length and consist of a spinel core enveloped by a Fe-rich silicate phase that probably is a (dehydrated?) phyllosilicate—too small to be analyzed with the electron microprobe. A few very small perovskite grains (<2 µm) are enclosed within the spinel. The chemical composition of the spinel is that of a Mg-Al spinel containing minor amounts of SiO_2 (0.21 wt%), TiO_2 (0.09%), Cr_2O_3 (0.11%), and FeO (0.83%). On top of the Fe-rich silicate envelopes there is a discontinuous rim of aluminous Ca-rich pyroxene with a fairly high FeO content (8.0 wt%). The trace-element content as determined by secondary ion mass spectrometry (SIMS) of these inclusions resembles that of group II CAIs [2].

Meanwhile we have found a second Antartic micrometeorite containing a few spinel grains. This spinel is associated with some tiny ilmenite grains and embedded in the foamy melt matrix of scoriaceous micrometeorite particle MM94/1-28. This particle is about 80 µm long and consists mainly of a highly vesicular melt of chondritic composition, which encloses the spinel grains (up to 10 µm long) and a few particles of thermally altered former phyllosilicates. The chemical composition of the spinel is that of a Mg-Al-spinel containing small amounts of FeO (0.6 wt%), but no Cr_2O_3. We have successfully analyzed the O isotopic composition of two spinels from MM92/15-23 and one from MM94/1-28 following the procedures as outlined by [3]. The O isotopic compositions were found to be ($\delta^{17}O/\delta^{18}O$ relative to SMOW in ‰)–25/–23, –24/–14, and –23/–24 in spinel 92/15-23/1 and 2 and 94/1–28 respectively. The data are plotted in Fig. 1 with the approximate 2σ error bars of ±8‰. Two of our data points plot directly onto the Allende mixing line [e.g., 4] approximately halfway between the spinels richest in ^{16}O and the terrestrial fractionation line. Such an isotopic composition of O is typical for spinel-rich type II inclusions [e.g., 5]. One spinel of MM92/15-23 plots to the right of the Allende mixing line. However, considering the analytical uncertainties, its O isotopic composition is compatible with the Allende mixing line. All spinels are much less ^{16}O-rich than a CAI found among IDPs [6].

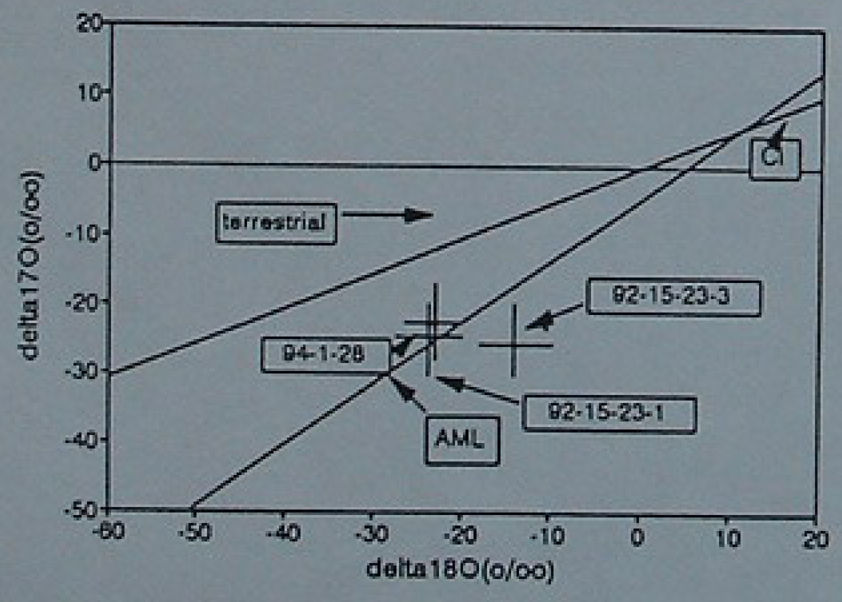

Fig. 1. Isotopic composition of O in spinels from Antarctic micrometeorites. AML: Allende mixing line.

In conclusion, the most common matter accreting onto the Earth today and represented by unmelted and partially melted micrometeorites consists of a matter similar, but not identical, to CM carbonaceous chondrites [e.g., 7]. The presence of spinel-rich CAIs with trace-element contents and O isotopic compositions of group II inclusions provides an additional support of that view.

References: [1] Kurat et al. (1994) *LPS XXV*, 763–764. [2] Martin P. M. and Mason B. (1974) *Nature, 249*, 333–334. [3] Zinner E. (1989) *USGS Bull., 1890*, 145–162. [4] Clayton R. N. (1993) *Annu. Rev. EPS, 21*, 115–149. [5] Clayton R. N. et al. (1986) *LPS XVII*, 139–140. [6] Stadermann F. J. (1991) *LPS XXII*, 1311–1312. [7] Kurat et al (1994) *GCA*, in press.

GEOCHEMISTRY AND METALLOGRAPHY OF THE CHLUMEC IRON: A METEORITE?

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A peculiar iron was recently found in Chlumec nad Cidlinou, Czech Republic [1]. It is rich in silicate inclusions and contains some C-rich inclusions. There are abundantly present large, irregularly shaped voids. The silicate and C inclusions are described by [1]. We report on the bulk metal composition and the metal texture.

The metal is fine-grained granular (~150 µm) with some primary lamellae (1–10 µm wide) and very abundant secondary deformation twins (Neumann lines). The Vickers hardness is between 362 and 567 kp/mm² (indentation diagonal dimension 127–160 µm). An X-ray analysis shows kamacite and taenite in about equal proportions. Silicate and sulfide inclusions are arranged in strings and swirls resembling the texture of the Tucson iron [e.g., 2].

Analysis of the metal with an electron microprobe revealed a fairly homogeneous composition with an average Ni content of 9.3 wt% (σ = 0.12) and a Ni/Co ratio of 37. The metal contains Cr but little Si and P (<0.03 wt%). Troilite is always Cr-bearing with highly variable Cr-contents ranging from 1 to 7.5 wt%. The Ni content ranges from 0.3 (Cr rich) to 1.5 wt% (Cr poor).

Bulk trace-element contents of a metal-rich sample as determined by INAA are plotted in Fig. 1. The refractory siderophile elements are strongly fractionated, as are the volatile elements. Chlumec does not fit any known Fe meteorite group [e.g., 5].

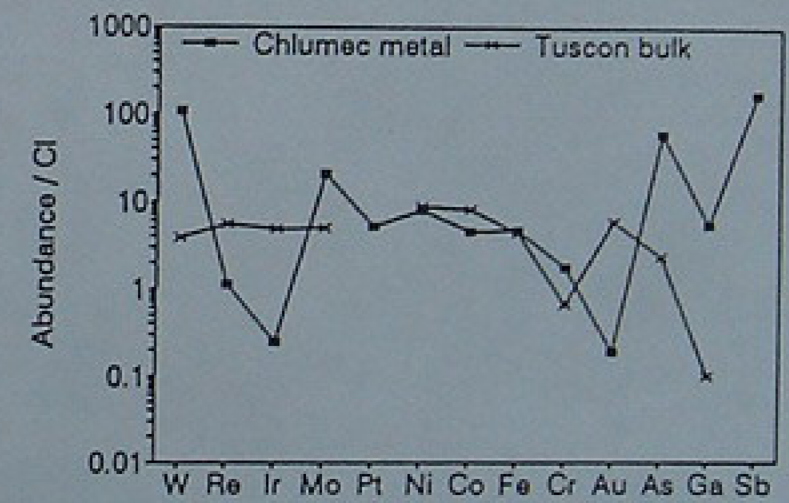

Fig. 1. Orgueil-normalized [3] major- and trace-element contents of Chlumec compared to those of Tucson [4].

Chlumec certainly has a texture untypical for either Fe meteorite or technical product. The original granular texture has been mechanically deformed. The Chlumec metal could also have been exposed to shock. It does not resemble any modern steel known to us.

The Cr contents of the metal and the troilite indicate fairly reducing conditions, which is contradicted by the presence of Fe-rich silicates [1]. The fractionation of the refractory siderophile elements is similar to but more extreme than that observed in E chondrites [e.g., 6] and Kaidun metal [7] and indicates an oxidized precursor. The high contents of Sb, As, Mo, and W could indicate a terrestrial origin. On the other hand, the presence of Ir and high contents of Pt as well as the only slightly fractionated Ni/Co ratio seem to point toward an extraterrestrial origin. The lack of any cosmogenic nuclides in Chlumec [8] indicates either perfect shielding (in a large body) an old terrestrial age, or a terrestrial origin. Continuing studies should help solve the puzzle of Chlumec.

References: [1] Bukovanská M. et al., this volume. [2] Nehru C. E. et al. (1982) *Proc. LPSC 13th*, in *JGR, 87*, A365–A373. [3] Palme H. et al. (1981) in *Landoldt-Boernstein* (K. Scheifers and H. H. Voigt, eds.), *2*,