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## Petrology and geochemistry of Antarctic micrometeorites

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**Abstract**—The petrology and geochemistry of twenty-three chondritic dust particles with masses of 1–47  $\mu\text{g}$  (sizes 100–400  $\mu\text{m}$ ) were recovered from blue ice near Cap Prudhomme, Antarctica, and studied by INAA, ASEM, EMPA, and optical microscopy. Sample selection criteria were irregular shape and (for a subsample) black color, with the aim of studying as many unmelted micrometeorites (MMs) as possible. Of thirteen unmelted MMs, six were phyllosilicate-dominated MMs, and seven were coarse-grained crystalline MMs consisting mainly of olivine and pyroxene. The remaining ten particles were largely melted and consisted of a foamy melt with variable amounts of relic phases (scoriaceous MMs). Thus, of the black particles selected, an astonishing portion, 40% (by number), consisted of largely unmelted MMs.

Although unmelted, most phyllosilicate MMs have been thermally metamorphosed to a degree that most of the phyllosilicates were destroyed, but not melted. The original preterrestrial mineralogy is occasionally preserved and consists of serpentine-like phyllosilicates with variable amounts of cronstedtite, tochilinite-like oxides, olivine, and pyroxene. The crystalline MMs consist of olivine, low-Ca pyroxene, tochilinite-like oxides, and occasional Ni-poor metal. Relics in scoriaceous MMs consist of the same phases. Mineral compositions and the coexistence of phyllosilicates with anhydrous phases are typical of CM and CR-type carbonaceous chondrites. However, the olivine/pyroxene ratio ( $\sim 1$ ) and the lack of carbonates, sulfates, and of very Fe-poor, refractory element-rich olivines and pyroxenes sets the MMs apart from CM and CR chondrites.

The bulk chemistry of the phyllosilicate MMs is similar to that of CM chondrites. However, several elements are either depleted (Ca, Ni, S, less commonly Na, Mg, and Mn) or enriched (K, Fe, As, Br, Rb, Sb, and Au) in MMs as compared to CM chondrites. Similar depletions and enrichments are also found in the scoriaceous MMs. We suggest that the depletions are probably due to terrestrial leaching of sulfates and carbonates from unmelted MMs. The overabundance of some elements may also be due to processes acting during atmospheric passage such as the recondensation of meteoric vapors in the high atmosphere.

Most MMs are coated by magnetite of platy or octahedral habit, which is rich in Mg, Al, Si, Mn, and Ni. We interpret the magnetites to be products of recondensation processes in the high ( $>90$  km) atmosphere, which are, therefore, probably the first refractory aerominerals identified.

### INTRODUCTION

THE EARTH ACCRETES between 15,000–50,000 t/a of extraterrestrial matter (HUGHES, 1978; KYTE and WASSON, 1986; ESSER and TUREKIAN, 1988; LOVE and BROWNLEE, 1993). Most of it ( $\sim 95\%$ ) is dust, within the mass range from  $10^{-8}$  to  $10^{-2}$  g corresponding to a particle size-range of about 40 to 1500  $\mu\text{m}$  (HUGHES, 1978). Curiously, this is the matter of which we are most ignorant because it was not available for study in the laboratory until very recently. Large extraterrestrial particles (centimeter to meter in diameter) are fairly unimportant for the current accretionary infall, but are very well studied (although still not very well understood, e.g., WASSON, 1985; KURAT, 1988). Also, the smaller dust fraction, between 5–15  $\mu\text{m}$ , has been fairly well characterized since samples became available some 20 years ago (“interplanetary dust particles”, IDPs, collected by aeroplanes in the stratosphere, e.g., BROWNLEE, 1981, 1985). The molten remnants of larger sized interplanetary dust, the “cosmic spherules,” which have been studied for over a century (and are still occasionally mistaken for large meteoroid ablation spherules), have also been fairly well characterized.

Unmelted dust of the main size-population has not been available for study for a number of reasons: First, theory predicts that dust particles  $>50$   $\mu\text{m}$  in size cannot survive atmospheric entry unaltered (KORNBLUM, 1969; HUGHES, 1978; LOVE and BROWNLEE, 1991). They are supposed to melt and partly evaporate due to frictional heating. Second, the rich mines of small extraterrestrial matter, the ocean floor sediments (BROWNLEE, 1985) and the Greenland cryoconite (MAURETTE et al., 1986), did not provide thermally unaltered extraterrestrial particles. However, the lack of such particles is likely to be an artefact. Worms and rhizopods (in the deep-sea sediments) and siderobacteria (in the cryoconite) are probably the culprits. Unmelted interplanetary dust is mostly friable and easily destroyed if attempts are made to metabolize it (by animals) or to free it from the cryoconite (by man). In any case, the nonavailability of the most common matter accreting onto the Earth has been accepted by the scientific community as being natural and unavoidable, and appropriate efforts were put into the collection of the small dust fraction from the stratosphere.

The presence of a fair amount of large, partly melted particles in cryoconite (MAURETTE et al., 1986, 1987; ROBIN et

al., 1990) led one of us (M. Maurette) to believe that there must be reservoirs of unmelted samples on Earth. The places where such particles would be the least exposed to terrestrial destructive processes are the polar ice sheets. The first attempt to mine such matter from the Antarctic ice at Cap Prudhomme near the French station Dumont d'Urville was highly successful. More than 20,000 particles (50–400  $\mu\text{m}$ ) were collected by melting about 100 t of ice (MAURETTE et al., 1989a,b, 1991). The largest part of this collection comprises partly melted and unmelted micrometeorites. We prefer to call the large unmelted dust particles "micrometeorites" in order to distinguish them from the much smaller IDPs or stratospheric dust particles. Thus, the first collection of micrometeorites outweighed the IDP collection, put together over 20 years (BROWNLEE, 1985), by a factor of about 1000. During the second expedition in the Antarctic summer 1990/91, 250 t of ice were melted and 20,000 more particles were added to the micrometeorite collection (MAURETTE et al., 1992a).

First studies of selected samples of this newly available, most important, and totally unknown, extraterrestrial matter showed that the unmelted micrometeorites

- 1) have approximately chondritic abundances of most refractory elements (with some exceptions) (MAURETTE et al., 1991; KOEBERL et al., 1992; KURAT et al., 1992a, 1993; PRESPEL et al., 1993);
- 2) are depleted in Na, Ca, Ni, S, Se, and Zn compared to CI chondrites and enriched in K, Rb, U, As, and Sb (KOEBERL et al., 1992; KURAT et al., 1992a,b; KLÖCK et al., 1992a; MAURETTE et al., 1991, 1992b; BECKERLING et al., 1992; PRESPEL et al., 1993);
- 3) are commonly enriched in C as compared to Orgueil (CI), and some carry a peculiar Fe-oxide rich in C, P, S, and other elements (COPS, see PERREAU et al., 1992, 1993; ENGRAND et al., 1993);
- 4) consist of hydrous and anhydrous silicates mixed in all possible proportions, and contain oxides, sulfides, and metal—comparable to the mineralogy of CM chondrites (BRANDSTÄTTER et al., 1991; CHRISTOPHE MICHEL-LEVY and BOUROT-DENISE, 1992; KLÖCK et al., 1992b; KURAT et al., 1992b, PRESPEL et al., 1992); and
- 5) contain solar energetic particle (SEP) gases, confirming that they have been exposed to the solar wind as micrometeoroids (MAURETTE et al., 1989c, 1991).

The search for isotopic anomalies and interstellar grains has so far been negative (ALEXANDER et al., 1992), but only a very small fraction of the available samples has yet been analyzed.

Here we report the results of a combined geochemical and petrological study of a suite of large micrometeorites selected from the 1991 EUROMET expedition to Antarctica (MAURETTE et al., 1992a). In this first study we concentrate on chondritic micrometeorites, which were purposely selected from the dust collection, which undoubtedly hosts a variety of dust objects, and probably also some nonchondritic extraterrestrial dust. Thus, this report should be understood as an attempt to characterize the chondritic portion of the interplanetary dust collected by the Antarctic ice.

Preliminary results have previously been reported by BRANDSTÄTTER et al. (1991), KOEBERL et al. (1992), KURAT et al. (1992a,b, 1993), MAURETTE et al. (1992b, 1993), PRESPEL (1993), and PRESPEL et al. (1992, 1993).

## DEFINITIONS

As indicated above we will call the large-sized unmelted interplanetary dust particle micrometeorites (MMs) (Figs. 1, 2). Such MMs may have experienced a variety of alterations, mainly by the action of frictional heat during atmospheric entry. Thus, we distinguish MMs without any sign of thermal alteration, those showing different degrees of thermal metamorphism, and those which were partly melted to various degrees. The latter grade, with increasing degree of melting, into scoriaceous micrometeorites (Fig. 3), which are quenched foamy melts with various amounts of relic phases. Depending on the peak temperature experienced and the duration of the thermal event, scoriaceous MMs are successively degassed and surface tension tends to shape them into droplets. These have been called cosmic spherules for a long time. This name applies to all round or droplet-shaped objects, regardless of the amount of vesicles they contain. Again, we see a gradual change from highly vesicular, scoriaceous cosmic spherules to the totally degassed, dense, and quench-textured chondritic cosmic spherules. Magnetite cosmic spherules are obviously a different variety.

Unmelted (and partly melted) interplanetary dust particles collected in the stratosphere have been dubbed interplanetary dust particles (IDPs). It is unfortunate that the general name for the total mass spectrum of dust has been given to the small tail ( $\sim 5\text{--}15\ \mu\text{m}$ ) of that dust size distribution, which, in addition to being small, contributes little to the total terrestrial infall (e.g., LOVE and BROWNLEE, 1993). Eventually, the name should be changed—and we strongly recommend to use the term stratospheric interplanetary dust particles (SIDPs, which could also stand for "small interplanetary dust particles").

Small particles (in the size range of SIDPs) must also be present in the Antarctic ice. However, they have not yet been collected. An attempt during the 1993/1994 field season (MAURETTE et al., 1992a) was, unfortunately, unsuccessful. Processing of ice samples should produce a collection of small dust samples in the near future.

## SAMPLES AND METHODS

From about 260 t of melt ice water, approximately 50 g of sediments were extracted in four size fractions: 25–50  $\mu\text{m}$ , 50–100  $\mu\text{m}$ , 100–400  $\mu\text{m}$ , and >400  $\mu\text{m}$  (MAURETTE et al., 1992a). The twenty-five daily collections during January 1991 are of different quality but, on average, contain a high proportion of extraterrestrial matter (about 10% of the particles <100  $\mu\text{m}$ ). About 80% of those, especially in the 50–100  $\mu\text{m}$  size-fraction, consist of partly melted to unmelted micrometeorites. The >100  $\mu\text{m}$  size-fractions contain a higher proportion of terrestrial debris, as well as cosmic spherules and scoriaceous MMs, than the smaller size fractions, but also contain unmelted micrometeorites. We selected particles from the 100–400  $\mu\text{m}$  size fraction, collected on January 15, 1991, for this study (samples 910115A and 910115B). Individual particles were handpicked under a binocular microscope. Selection criteria were irregular shape and dark color.

For instrumental neutron activation analysis (INAA), four sets of particles were selected, comprising seven to twelve particles each. These sets were labelled M, AM, 3M, and 4M. Each particle is identified by its number within a set following the set name (e.g., M1, M2, etc.). The samples were weighed with a Mettler UM3 ultramicrobalance. Sample masses ranged from 1–17  $\mu\text{g}$ , with one sample having 47  $\mu\text{g}$ . Particles were individually placed into small depressions on circular high-purity quartz disks. The disks were then stacked together with disks containing standards, which were placed on the top and bottom of the stack. Granite G-2 (USGS, GOVINDARAJU, 1984) and Allende meteorite powder (U.S. National Museum, JAROSEWICH et al., 1987) were used as standards. Two sets of both standards, weighing up to 1 mg each, were used for each irradiation.

The stack was packed in aluminum foil and irradiated in aluminum rabbits for about 120 h at the 10 MW ASTRA research reactor of the Forschungszentrum Seibersdorf, Niederösterreich, at a flux of  $6 \times 10^{13} \text{ n cm}^{-2} \text{ s}^{-1}$ .

After cooling for 2 days, samples and standards were transported to the counting laboratory at the University of Vienna, reweighed, and individually transferred into polyethylene capsules. The samples were counted utilizing HpGe detectors with up to 48% relative efficiency and 1.85 keV resolution at 1332 keV (see KOEBERL, 1993). This setup allows the determination of the concentrations of twenty-five to thirty-five elements in samples larger than about 2  $\mu\text{g}$  mass. Three measurement cycles were performed, with the first starting 2 days after the end of the irradiation and the last one about 1 month later. Counting times ranged from about 5 h to 3 days. The precision is believed to range between 10–20% for most elements. For standard deviations exceeding  $\pm 40\%$  only upper limits are given. The major factor affecting accuracy is the error made in weighing samples in the 1–10  $\mu\text{g}$  range. After cooling for several months, the particles were mounted on a glass slide, carbon coated, and studied with a JEOL6400 analytical scanning electron microscope. Subsequently, the samples were mounted in epoxy and polished. The sections were studied under an optical microscope and with an analytical scanning electron microscope. Bulk and mineral analyses were made with an automated ARL-SEMQ electron microprobe (EMP) operated at 15 kV acceleration potential and 15 nA beam current. Measurements were made against mineral standards. Standard Bence-Albee corrections were applied for matrix effects. Bulk rock analyses were made while scanning the electron beam (in TV mode) over areas poor in voids. No correction for matrix effects was applied. Bulk glass analyses were made by applying the mineral analysis procedure.

## RESULTS

A total of thirty-seven mostly black particles, but including a few that are transparent as well as some that are rusty, were analyzed by INAA. A few particles were lost during handling. Of the remaining particles, twenty-five turned out to be of approximately chondritic composition. Two of these were cosmic spherules and were not included in this dataset. Thus, twenty-three particles qualified to be included into this report, representing a total mass of 189.5  $\mu\text{g}$ .

### Mineralogy and Petrology

The twenty-three particles exhibit (as expected) a variety of textures. According to their texture and mineralogy (Appendix), they can be grouped into the following classes:

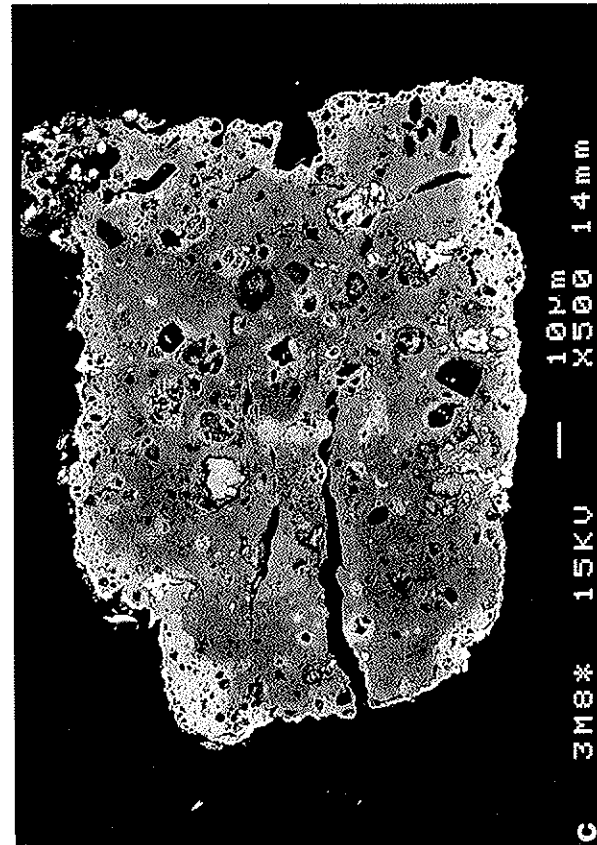
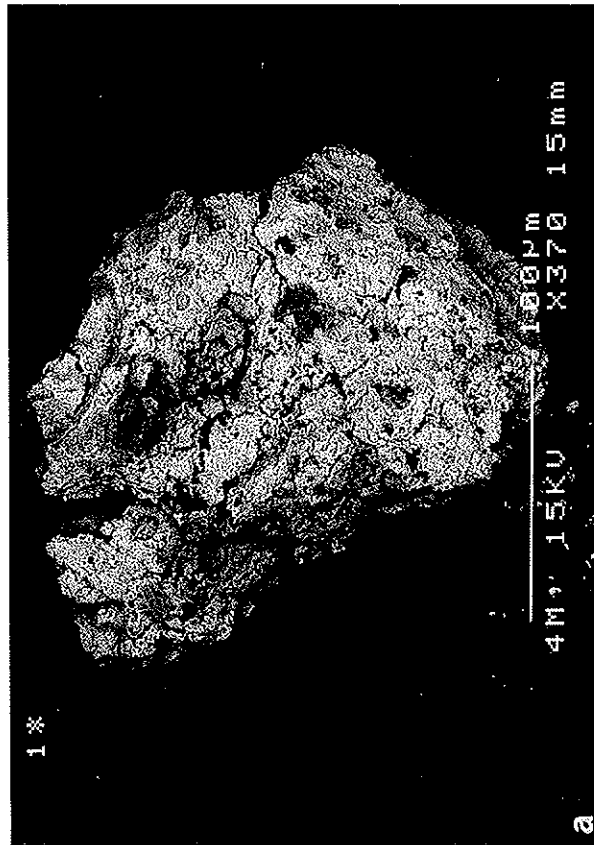
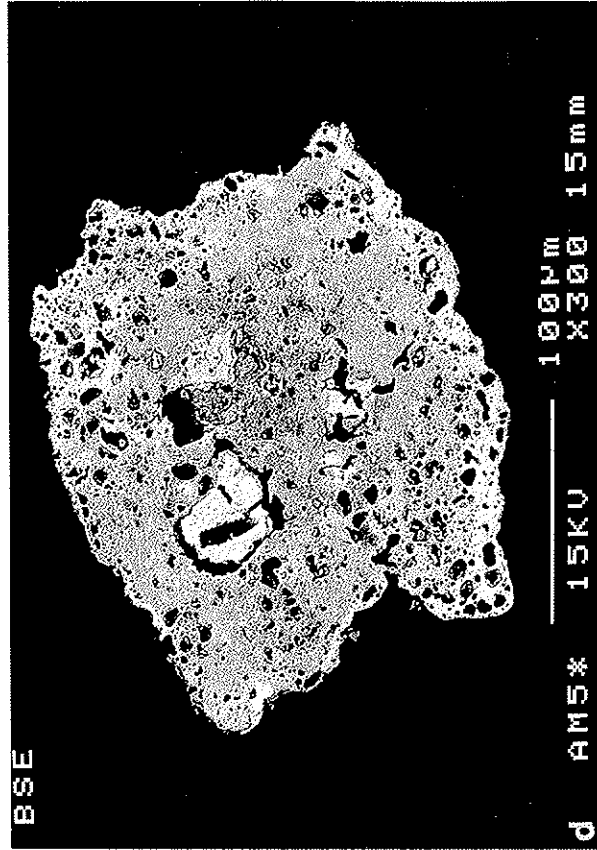
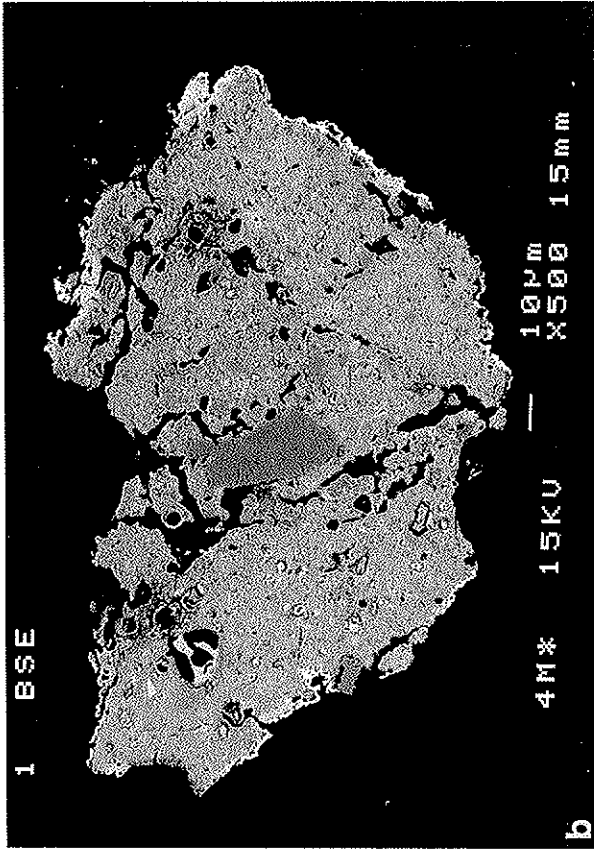
- 1) Largely unmelted micrometeorites (MMs) consisting either mainly of (a) phyllosilicates ("phyllosilicate MM") or of (b) coarse-grained anhydrous minerals ("crystalline MMs"), or of mixtures thereof (Figs. 1, 2).
- 2) Scoriaceous micrometeorites consisting of a highly vesicular melt (Fig. 3) which may contain various amounts of relic phases (mainly anhydrous silicates). Scoriaceous MMs grade, with increasing degree of degassing and re-shaping, into droplets (Fig. 3d).
- 3) Cosmic spherules. Transitional types can be either rich in vesicles or in relic anhydrous minerals. The typical final product is a compact, quench-textured, silicate-rich spherule.

Phyllosilicate MMs consist mainly of phyllosilicates, oxides, various amounts of pyroxenes, and olivines, and usually contain only trace amounts of sulfides. There is some variety of phyllosilicates present, but their nature cannot be estab-

lished properly by the methods available to us. Matrix phyllosilicates have typical chondritic major and minor element abundances (Table 1, AM4, 4M1), with only a few exceptions. Totals of the analyses vary, indicating different degrees of dehydration by thermal metamorphism during atmospheric entry (compare AKAI, 1988). Cronstedtite-like phases are quite common (Table 1, AM5). Oxides have complex compositions dominated by iron oxide, contain all major chondritic elements, and have low totals. This may indicate a high degree of oxidation and/or some volatile content. Sulfur is always present in minor amounts ( $\sim 0.1$ –1 wt%, not shown in Table 1). Some similarity to tochilinite is apparent. A laihunite-like phase was encountered in particle 3M8.

Coarse-grained crystalline MMs have mostly porphyritic textures and consist of olivine, pyroxene, plagioclase, and some not well-defined iron oxides set into a fine-grained or glassy matrix. Accessory phases are chromite, metal, and sulfides. Olivines range in composition from Fa 2 to Fa 37 and have highly variable minor element contents (Table 2). The low-Fe olivine of M1 has a high  $\text{TiO}_2$  content, intermediate contents of  $\text{Al}_2\text{O}_3$ ,  $\text{Cr}_2\text{O}_3$ , MnO, and CaO, and a FeO/MnO ratio of 15. Fayalite-rich (Fa 11–37) olivines are commonly poor in minor elements, with the exception of CaO and NiO, which can be high (M7, AM8, AM9). The FeO/MnO ratios vary from 24 to 45 for the low-Fe olivines, and from 60 to 130 for the Fe-rich ones. In many cases olivine compositions are variable within narrow limits. The low-Ca pyroxenes are FeO-bearing (11–13 wt%) and are either rich (M6) or poor (M1) in minor elements. Their FeO/MnO ratios are small (24–28). An augite was encountered in one sample (M1). Plagioclase (An 12) is present in two particles (M7 and AM9) where it coexists with Fa-rich olivine (Fa 30 and Fa 35, respectively), one of which is also rich in NiO (0.5 wt%, M7) and coexists with Cr- and Ni-bearing magnetite. Apparently hydrated oxides are present in AM8. They are rich in minor elements and particularly rich in Al and Cr. These oxides resemble low-S tochilinite. Chromites or Cr-rich spinels are common, but usually too small for analysis ( $< 5 \mu\text{m}$ ). A Cr-poor Al-spinel was found in particle AM8, coexisting with Ni-rich olivine and tochilinite-like oxide.

Scoriaceous MMs are naturally dominated by glass, quench olivine, and quench magnetite—all too fine-grained to allow analysis. Scoriaceous MMs commonly contain relic phases which comprise olivine, pyroxene, plagioclase, iron oxides, and phyllosilicates (Table 3). Most particles contain Fe-poor olivines (Fa 1–2), which are always rich in Cr and Mn, resulting in an FeO/MnO ratio of 3–12. Only one Fe-poor olivine contains larger amounts of Ti, Al, and Ca (M4). Fayalite-rich olivines range from Fa 23 to Fa 31. This range is covered by the olivine of particle M2 which is unusual in being exceptionally rich in  $\text{TiO}_2$  (0.59 wt%),  $\text{Al}_2\text{O}_3$  (2.69),  $\text{Cr}_2\text{O}_3$  (0.36), NiO (1.07), and CaO (0.75), and which seems to contain some alkali elements. A second Fa-rich olivine (Fa 28) is present in particle M4, where it coexists with a forsterite rich in minor elements. This Fa 28 olivine also contains Ni, some Cr, and Al. The FeO/MnO ratios of fayalitic olivines are 87 and 132, respectively. Pyroxenes are typically enstatites (Fs 1–5), which are rich in minor elements (Ti, Al, Cr, Ca) and sometimes contain NiO (about 0.1 wt%).



Their FeO/MnO ratios are all very low (4–12). One augite grain was found in particle 3M2 (coexisting with enstatite, plagioclase, and iron oxide) which is rich in TiO<sub>2</sub> (0.45 wt%), Al<sub>2</sub>O<sub>3</sub> (4.4%), Cr<sub>2</sub>O<sub>3</sub> (2.1%), and particularly in MnO (2.68%). Its FeO/MnO ratio is only 0.75. Plagioclase (An 25) coexisting with this unusual augite in particle 3M2 is also unusually rich in minor elements such as Ti, Fe, Mn, and Mg, and has a FeO/MnO ratio of about unity. Iron oxides are common but have diverse compositions. All of them have high NiO contents (1–1.4 wt%), and MgO contents ranging up to 11.2 wt% suggest a solid solution with Mg(OH)<sub>2</sub>.

One scoriaceous MM (4M12) has preserved a core of phyllosilicates of chondritic composition. The analysis shows depletions in Na, Ni, Ca, and Mn, a feature commonly found in phyllosilicate MMs (compare KURAT *et al.*, 1992a,b). Phyllosilicates in 4M12, however, appear to be thermally metamorphosed, as indicated by the high total of the analysis and the lack of birefringence.

### Bulk Compositions

Bulk major and minor element contents of micrometeorites, as determined by broad-beam electron microprobe techniques, are given in Table 4. Bulk data acquired by INAA are given in Table 5. The last column in Table 5 shows the range of precision for the determination of the elemental concentrations. The precision varies somewhat for individual analyses, because of factors such as difference in sample mass, elemental concentration, counting parameters, and spectral interferences. Bulk data acquired by EMPA techniques represent areas of a given object, which could be probed by the electron beam. In many cases the size of these areas is restricted and appropriate spots were usually found in the center of the particles. It is therefore not surprising that bulk abundances determined by EMPA and INAA occasionally disagree. The most common case shows considerably higher abundances for the real bulk analysis, as determined by INAA, as compared to EMPA data. This holds for siderophile (Fe, Ni) and lithophile (Cr, Na) elements (Fig. 4). These elements appear to be inhomogeneously distributed within a given particle, with enrichments preferentially at or near the surface. The enrichment of Fe and Ni is commonly visible as magnetite envelopes (see below). Occasionally Na (as NaCl) is also present at the surface.

### Major and minor elements

Major and minor element abundances in phyllosilicate MMs are mostly chondritic, but some fractionations are ap-

parent (Fig. 5). The phyllosilicate MMs tend to be depleted in Ca, Na, Ni, and S (Fig. 5a) as compared to CI chondrites. Potassium can be either depleted or enriched. Crystalline MMs naturally show a wider spread in major and minor element abundances (Fig. 5b) and seem to form two compositional groups: One group shows unfractionated abundances of lithophile elements Al, Ti, Ca, Mg, Si, Mn, and Na, and depletions in Cr and K; these samples comprise porphyritic fragments. The second group, consisting of an olivine crystal (M7) and a porphyritic fragment, is depleted in Al, Ti, Ca, Cr, Na, and K. Both groups show strong depletions in Ni and S and have either high (second group) or low (first group) Fe contents. Scoriaceous MMs have major and minor element abundances similar to those of phyllosilicate MMs (Fig. 5c), but with more variation in the abundances and tendencies towards strong depletions in Na and K contents. However, when measured, the Na/K ratio is always below that of CI chondrites and one sample (3M5) is even enriched in K compared to CI. Also, Ni and S are depleted, with Ni approximately to the same level as in phyllosilicate MMs, but S to much lower levels.

### Trace elements

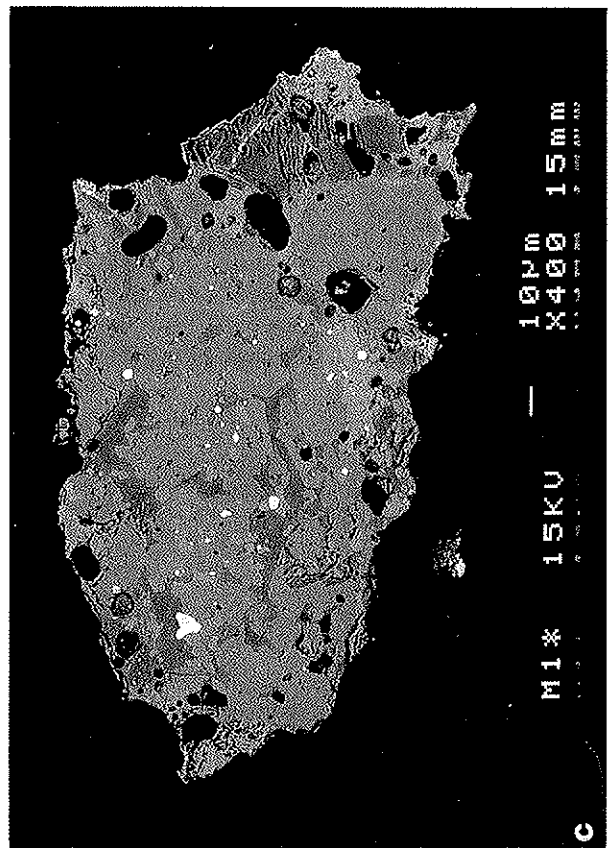
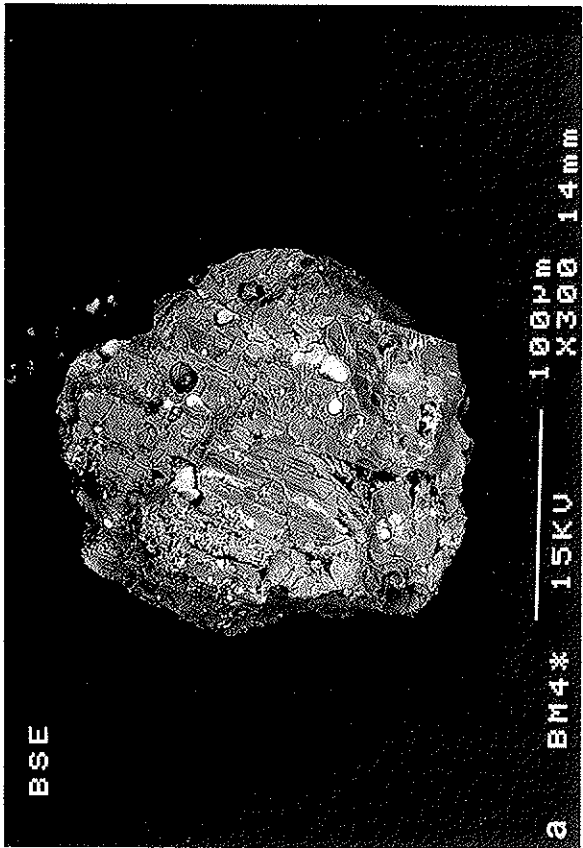
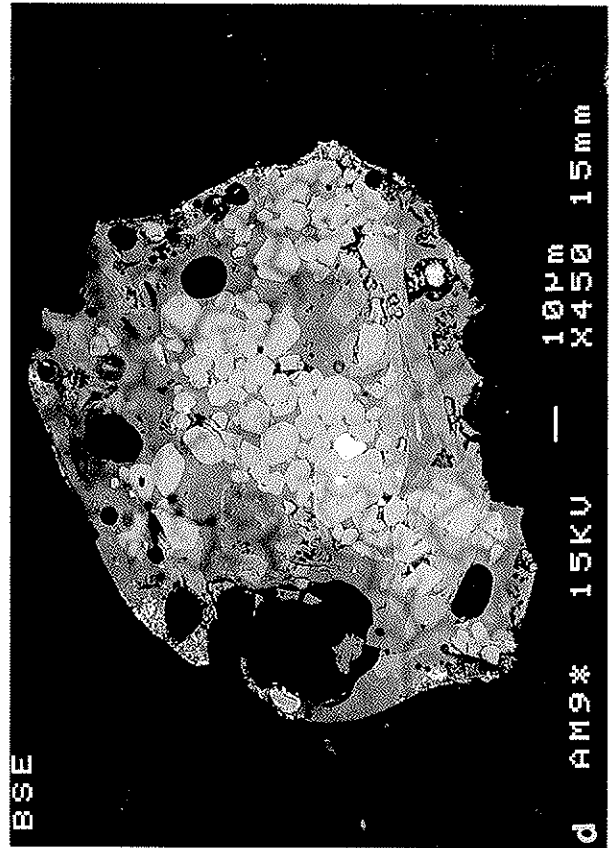
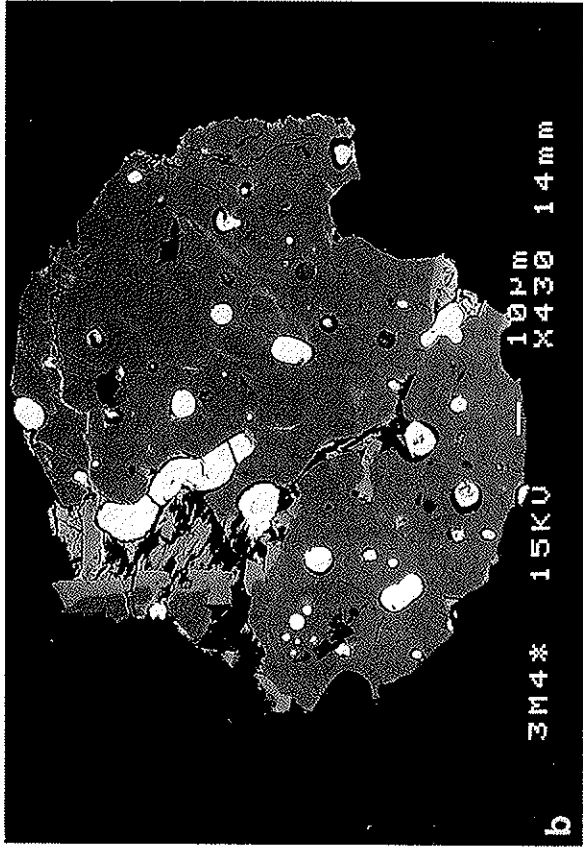
Trace element abundances in phyllosilicate MMs are mainly chondritic. Refractory lithophile elements (Fig. 6a) are, in general, unfractionated, as is the abundance of Cr. Sodium and Zn are depleted relative to CI and K and Br are often enriched ( $0.7\text{--}15 \times \text{CI}$  for Br).

The crystalline MMs show some more dispersion in lithophile element abundances (Fig. 6b). The refractory lithophile elements are enriched and fairly unfractionated. Chromium, and the more volatile elements, show large variations and range from depleted to enriched, except for Zn, which seems to be generally depleted ( $0.2\text{--}0.3 \times \text{CI}$ ).

The scoriaceous MMs show refractory lithophile element patterns similar to those of the other MMs (Fig. 6c). However, some show strong fractionations of Sc from the REEs, which is also seen in other micrometeorites, but less pronounced. Chromium abundances are variable but not as much as among the crystalline MMs. The volatile lithophile elements are generally depleted but their abundances are highly dispersed and range from highly depleted to enriched (e.g.,  $0.02\text{--}1.1 \times \text{CI}$  for Na,  $0.06\text{--}2 \times \text{CI}$  for Zn, and  $0.2\text{--}12 \times \text{CI}$  for Br). Again, K contents are high compared to other elements of comparable volatility.

Siderophile elements in phyllosilicate-rich MMs show a peculiar pattern (Fig. 7a). The refractory siderophile elements (Os, Ir) have unfractionated chondritic abundances. The

FIG. 1. Backscattered electron (BSE) scanning microphotographs of phyllosilicate-dominated micrometeorites. (a) Particle 4M1, almost totally covered by magnetite and cut by cracks. (b) Polished section of micrometeorite 4M1. Fine-grained (probably dehydrated) phyllosilicates and oxides with a large enstatite (Fs 1.5) in the center and a small one at the surface (left). The particle has several irregular voids, is cut by cracks, and is partly covered by magnetite (light-grey to white). (c) Polished section of micrometeorite 3M8 consisting of densely intergrown fine-grained (probably largely dehydrated) phyllosilicates and a laihunite-like phase (lighter grey) and almost totally covered by a scoriaceous, Fe-rich mantle (light grey) which in turn is covered by magnetite (white). The particle has abundant irregular voids and some open cracks. (d) Polished section of micrometeorite AM5 consisting of fine-grained phyllosilicates including cronstedtite (light grey) and tochilinite (white). The particle has abundant irregular voids, is enveloped by a scoriaceous mantle of variable thickness, and covered by a thin magnetite crust.



moderately volatile elements Ni and Co are depleted relative to Ir, with Ni being more strongly depleted than Co. Iron is enriched over CI abundances and Au, As, and Sb are strongly enriched (up to  $50 \times$  CI). Selenium is slightly depleted ( $0.3\text{--}0.8 \times$  CI).

The crystalline MMs display considerably more spread in their siderophile element contents. On average, they are depleted in Os, Ir, Ni, Co, Au, Sb, and Se as compared to CI chondrites (Fig. 7b). Os/Ir and Ir/Ni ratios are mostly unfractionated but Co/Ni ratios are commonly fractionated to  $\text{Co/Ni} >$  or  $<$  CI. An exception is particle AM1, which has  $\text{Ir/Ni} \gg$  CI. Only one crystalline MM is rich in siderophile elements (3M4) with Os, Ir, Fe, Au, As, Sb  $>$  CI  $\sim$  Ni, Co  $>$  Se, and a peculiar Os/Ir ratio. Most crystalline MMs show volatile siderophile element patterns comparable to those of phyllosilicate MMs. However, Au is depleted relative to Fe in the crystalline MMs in contrast to the phyllosilicate MMs, where Au is enriched over Fe. In most cases the Ir/Au ratio is lower than that of CI chondrites, similar to that of phyllosilicate MMs.

Scoriaceous MMs (Fig. 7c) have siderophile element patterns that are very similar to those of phyllosilicate MMs. Minor differences are a larger spread in abundances, a smaller Ni depletion (relative to Ir), and somewhat lower Au abundances in the scoriaceous MMs compared to the phyllosilicate MMs.

### Magnetite Envelopes

Most micrometeorites are partly to totally enveloped by magnetite or a magnetite-like phase (Figs. 1–3, 8). Envelopes, usually  $0.5$  to  $1 \mu\text{m}$  thick, but sometimes reaching  $4 \mu\text{m}$ , are present on all types of MMs, but are particularly common and well developed on scoriaceous MMs. Several generations of magnetites are commonly present with morphologies ranging from platy to octahedral and (in vesicles) acicular. In polished sections (Fig. 8d), the succession of different generations and the granular nature of the magnetite envelope is occasionally visible. Magnetite covers are commonly associated with Fe enrichment (metasomatism) of the substrate (Figs. 1, 3). The chemical composition of the surface magnetites (Table 6) is characterized by high contents of  $\text{SiO}_2$ ,  $\text{Al}_2\text{O}_3$ , MnO, MgO, and NiO.

### DISCUSSION

Despite the crude selection criteria, we were surprisingly successful in picking largely unmelted micrometeorites and, especially, phyllosilicate-dominated MMs from the  $100\text{--}400$

$\mu\text{m}$  dust fraction. Naturally, the scoriaceous MMs are the most common ones (ten out of twenty-three) followed by the physically resistant coarse-grained crystalline MMs (seven out of twenty-three). The high abundance of largely unmelted phyllosilicate MMs (26%) is surprising. It may reflect a high survival rate, despite the fact that the  $100\text{--}400 \mu\text{m}$  size fraction is rich in (totally melted) cosmic spherules ( $\sim 80\%$ ). Although the abundance of cosmic spherules of this dust size fraction is not known exactly, rough estimates indicate a spherule/nonspherule ratio of about 4 (MAURETTE et al., 1992a). This lowers the absolute abundance of unmelted dust to 10% and that of the phyllosilicate MMs to about 5%, which is higher than the most optimistic theoretical estimates (e.g., LOVE and BROWNLEE, 1991).

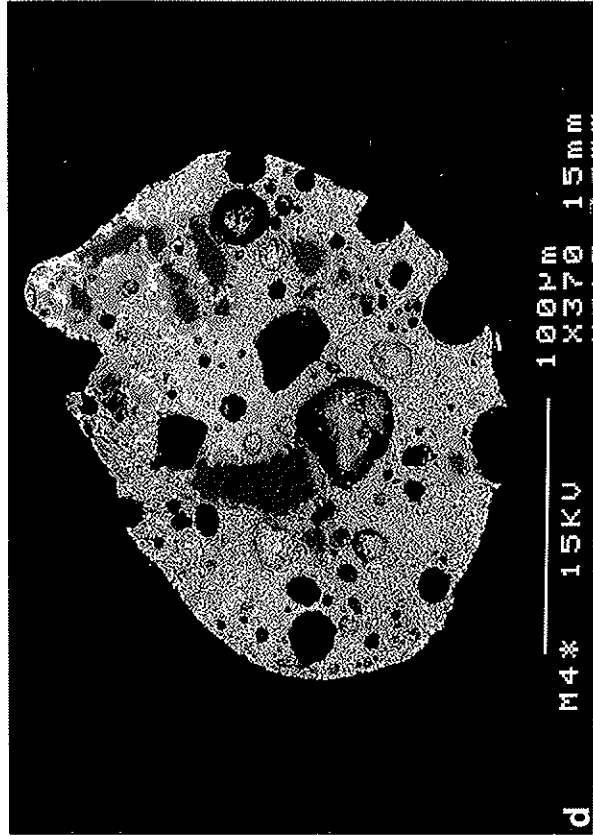
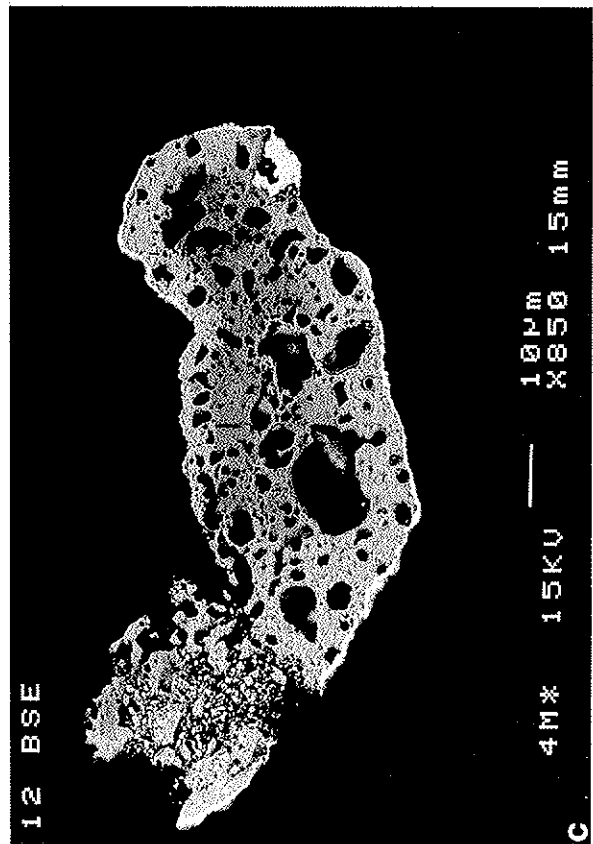
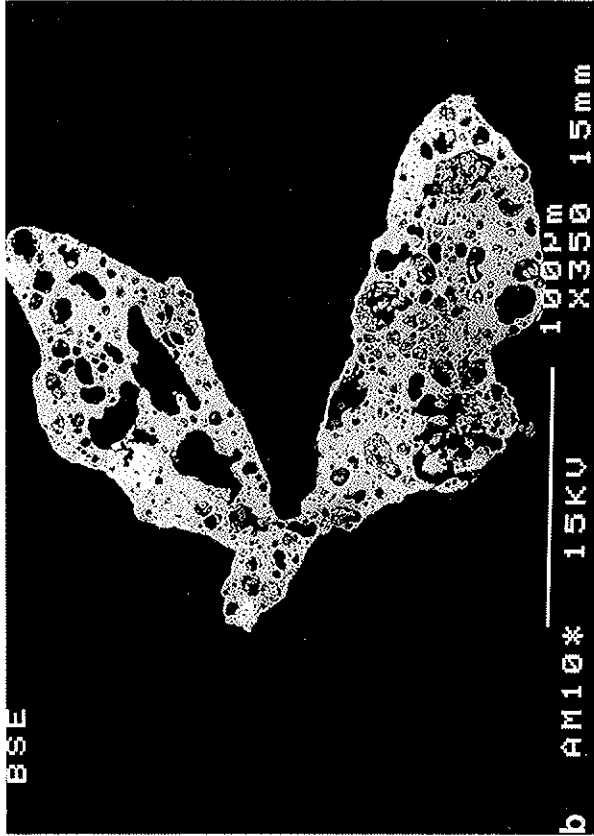
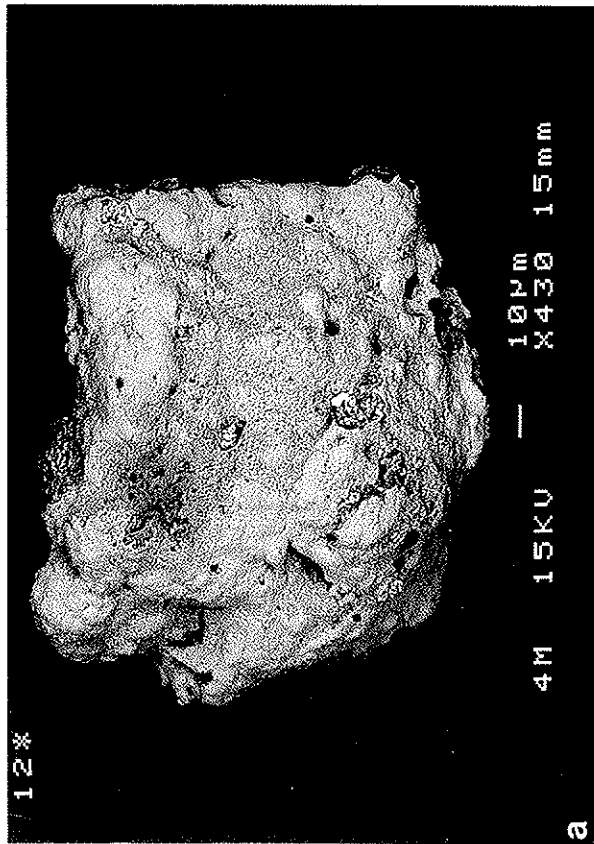
The number of crystalline MMs in our total sample is not representative because we deliberately picked some transparent particles. However, the proportion of phyllosilicate to scoriaceous MMs is significant because both types fit our selection criteria. The 1:1 proportion is certainly surprising considering theoretical predictions (KORNBLUM, 1969; HUGHES, 1978; LOVE and BROWNLEE, 1991).

The crystalline MMs comprise about 50% of the total mass ( $171.5 \mu\text{g}$ ); the phyllosilicate and scoriaceous MMs about 25% each. The average mass of an individual phyllosilicate MM is  $6.8 \mu\text{g}$ , that of a scoriaceous MM  $4.6 \mu\text{g}$ , and that of a crystalline MM  $12.1 \mu\text{g}$ . However, more than 50% of the total mass of crystalline MMs is represented by particle AM1 ( $47 \mu\text{g}$ ). Excluding this particle, the average mass of crystalline MMs becomes  $6.3 \mu\text{g}$ , comparable to that of the other micrometeorite types. All of them are on average more than 1000 times more massive than the average stratospheric dust particle (SIDP) of about  $10 \mu\text{m}$  diameter. However, the average mass of our micrometeorites is far from that of meteorites and even inferior to that of most meteorite components (e.g., chondrules). Several questions arise: Are micrometeorites representative samples of dust from the primitive solar nebula or do they represent components of larger units (e.g., chondrites); what are the similarities and differences between micrometeorites and meteorites and micrometeorites and SIDPs?

### Mineralogy and Mineral Chemistry

The mineralogy of unmelted micrometeorites is primarily determined by phyllosilicates, anhydrous Mg-Fe silicates, and mixtures thereof. Thus, at first glance, a relationship with CM chondrites is apparent. Phyllosilicate compositions (Table 1) indicate the presence of serpentine-like minerals and cronstedtite (AM5). These phases are typical of CM chon-

FIG. 2. BSE scanning microphotographs of coarse-grained crystalline micrometeorites. (a) Particle 3M4 consisting of a large enstatite which contains rounded inclusions of tochilinite and magnetite. The surface is slightly dusted by fine-grained magnetite. (b) Polished section of micrometeorite 3M4 consisting of a large enstatite crystal (Fs 3), poikilitically including small round olivines (Fa 2, dark grey) and magnetite (white) and large tochilinite (white). Fa-rich olivine is associated with tochilinite. A few very small magnetites (white) are present at the surface (right side). (c) Porphyritic micrometeorite M1, consisting mainly of orthopyroxene (Fs 20–23), clinopyroxene, chromite, sulfide, and metal (all white), and glass (dark grey). At the right side the porphyritic MM is intergrown with olivine of Fa 2–4 (dark grey). Note the abundant rounded voids. (d) Porphyritic micrometeorite AM9 consisting of olivine (Fa 32–37, light grey), augite (grey), plagioclase (An 12, dark grey), and glass (dark grey). Light phases are sulfide and chromite. The particle has abundant rounded voids and is covered by magnetite only in a few places.





drites (e.g., MÜLLER et al., 1979; ZOLENSKY and MCSWEEN, 1988; BROWNING et al., 1991; GRAHAM and KURAT, 1991). However, serpentine-saponite intergrowths, typical of CI chondrites, have also been described from a single MM (KELLER et al., 1992).

The anhydrous phases may, as a first approximation, be considered representative of one source sample, regardless of their occurrence (in phyllosilicate, crystalline, or scoriaceous MMs).

Olivines form two distinct compositional groups: a forsteritic one ( $Fa < 5\%$ ) and a more FeO-rich one ( $Fa > 10\%$ ). This is similar to the population of olivine in CM and CR chondrites (e.g., NELEN et al., 1975; GRADY et al., 1986; STEELE, 1986; KURAT et al., 1989a; WEISBERG et al., 1993). However, the low-Fe olivines of MMs are, on average, richer in FeO than those of CM and CR chondrites (also shown by larger numbers of samples; e.g., BRANDSTÄTTER et al., 1991; STEELE, 1992; PRESPEER et al., 1992) and poorer in refractory minor elements such as Al and Ca. They fall, however, within the compositional range shown by CM and CR forsterites. The same holds for elements of medium volatility, e.g., Cr and Mn, which are fairly abundant in MM forsterites. As a result, the FeO/MnO ratios of the MM forsterites are low (3–15), indicating a primitive heritage comparable to that of carbonaceous chondrite olivines.

The FeO contents of Fe-rich olivines range from 15.9 to 31.4 wt% (Tables 2 and 3). Most of these olivines represent individual grains which presumably have retained their original compositions. Only in particle M4 is there clear evidence for the in-situ formation of Fe-rich olivine from a forsteritic one, probably during melting upon atmospheric entry. This olivine is peculiar in being rich in NiO (0.77%) and in containing other minor elements (Al, Cr, Mn, Ca) at a level comparable to that of the original forsterite. This composition is apparently the result of the transformation of a forsterite rich in minor elements into a more fayalitic olivine under highly oxidizing conditions (as indicated by a high NiO content). The almost chondritic FeO/MnO ratio of 130 differs highly from that of the forsteritic olivine (9) and apparently reflects the bulk ratio.

The other FeO-rich olivines typically contain substantial amounts of MnO beside FeO resulting in FeO/MnO ratios between 40 and 130. Interestingly, the two olivines with the highest FeO/MnO ratio also contain appreciable amounts of NiO (about 0.5%). Only one of the most FeO-rich olivines (AM9) has an elevated CaO content. Thus, most FeO-bearing olivines in MMs appear to be similar in composition to those found in carbonaceous chondrites (e.g., HOINKES and KURAT, 1975; NELEN et al., 1975; KURAT et al., 1989a), but

differ in being poor in Cr, which is rare among CM and CR olivines, and in sometimes being rich in Ni. Olivines rich in Ni are known from the rare CK chondrites (KALLEMEYN et al., 1991; KURAT et al., 1991). There, olivine typically coexists with magnetite (Ni- and Cr-bearing) and intermediate plagioclase. This is exactly the mineral association found in crystalline MM M7 (Table 2), which could well represent a rock similar to CK chondrites. The second Ni-rich olivine encountered coexists with an oxide phase probably related to tochilinite (AM8, Table 2). This phase, however, is typical of CM chondrites. One olivine found in scoriaceous MM M2 (Table 3) has a highly peculiar composition in being extremely rich in  $TiO_2$  (0.59 wt%),  $Al_2O_3$  (2.69), NiO (1.07), and CaO (0.75) and in containing some  $Na_2O$  (0.19%). Such olivines are, so far, known only from a rock fragment from the Allende chondrite (All-AF, KURAT et al., 1989b; PALME et al., 1989). The texture of that rock fragment, constituting an aggregate of protochondrules, as well as the very high trace element contents of its olivines (KURAT et al., 1989c), make it the most primitive chondritic rock known. The olivine rich in minor elements in MM M2 is relict and no other primary silicate phase was found. However, its peculiar composition makes it likely that it was derived from a rock similar to All-AF. Thus, the chemical composition of olivines can be interpreted as indicating several sources: (1) CM and CR-like chondrites, (2) CK-like chondrites, and (3) an All-AF-like chondritic rock.

The notation "source chondrite" does not imply that MMs are physical derivatives of chondritic rocks. Rather, it intends to indicate that MMs come from a region in the solar nebula where constituents of a certain chondritic rock have been processed before accretion took place. This holds for all comparisons made in this paper and in no case implies that MMs have been derived from a certain chondritic rock by comminution, although such an origin cannot be excluded with certainty.

Pyroxene is approximately as abundant in the micrometeorites as olivine (see also the results derived from a larger sample of MMs by PRESPEER et al., 1992; KURAT et al., 1993). Such a high pyroxene abundance is unusual for CM chondrites, with the exception of Kaidun, a complex breccia of CM, CI, E, and other chondrites and achondrites (IVANOV, 1989; BRANDSTÄTTER et al., 1992), but is a characteristic feature of CR chondrites (e.g., WEISBERG et al., 1993). Low-Ca pyroxenes are the most common pyroxenes in MMs. They are of orthopyroxene composition ( $CaO = 0.26\text{--}0.58\%$ ), low in FeO (1.3–3%), and rich in minor elements (Ti, Al, Cr, Mn, and in some cases also Ni). Their FeO/MnO ratios are low (10–13). All features, except the occasional Ni contents,

FIG. 3. BSE scanning images of scoriaceous micrometeorites: (a) Particle 4M12 with rectangular shape, rounded corners, and dense magnetite coating. (b) Polished section of micrometeorite AM10 consisting of two highly vesicular club-shaped drops with a discontinuous magnetite cover. Note the different shades of grey which indicate variation in Fe contents. (c) Polished section of micrometeorite 4M12 (Fig. 3a) displaying a relictic phyllosilicate core (dark grey) enveloped by an Fe-enriched scoriaceous mantle (light grey) and covered by magnetite (white). The light phase at right is Fe-oxide. (d) Polished section of micrometeorite M4, a scoriaceous MM transitional to cosmic spherules. The almost droplet-shaped MM has a dense quench texture (silicate plus magnetite), abundant vesicles (some of rectangular shape), and contains abundant olivine (Fa 1, dark grey) which has partly reacted to form Fe-rich olivine (Fa 28). The particle is incompletely covered by magnetite.

TABLE 1: CHEMICAL COMPOSITION OF PHASES IN PHYLLOSILICATE-RICH MICROMETEORITES.

N	AM4		AM5		3M8		4M1		4M12							
	Matrix	Phyllo	Phyllo	Phyllo	Laihun	Phyllo	Pyrox	Matrix	Matrix	Phyllo*						
SiO <sub>2</sub>	30.6	27.2	38.3	24.8	33.1	15.5	12.3	42.7	6.90	4.40	24.3	51.4	56.6	25.2	32.7	40.9
TiO <sub>2</sub>	0.13	0.10	0.10	0.16	0.11	0.03	<0.02	0.12	0.06	0.05	0.07	0.09	0.13	0.12	0.12	0.07
Al <sub>2</sub> O <sub>3</sub>	2.30	4.30	2.82	1.97	2.00	0.81	0.87	3.20	0.98	1.50	1.88	3.90	0.79	1.51	2.24	3.00
Cr <sub>2</sub> O <sub>3</sub>	0.29	0.70	0.63	0.16	0.16	0.48	0.48	0.53	0.75	0.45	0.38	0.72	0.60	0.43	0.68	0.64
FeO	34.8	38.7	21.1	34.8	38.3	57.7	63.2	23.1	58.2	65.1	55.8	13.6	1.02	30.9	30.5	16.0
MnO	0.24	0.18	0.16	<0.02	<0.02	0.15	0.16	0.03	0.89	1.16	0.06	0.07	0.08	0.27	0.32	0.08
MgO	21.6	20.4	13.9	1.41	1.45	0.37	0.31	19.3	7.00	7.30	0.42	23.5	38.1	15.3	18.9	26.2
NiO	0.19	0.12	1.92	0.20	0.24	2.09	1.73	0.62	0.67	0.94	1.87	0.26	0.05	1.30	0.81	0.43
CaO	0.26	1.12	0.62	0.41	0.13	0.14	0.09	0.13	0.10	0.13	0.25	0.18	0.58	2.27	0.78	0.08
Na <sub>2</sub> O	0.19	0.10	0.82	0.85	0.54	2.27	1.92	0.80	1.18	0.63	0.44	0.35	<0.02	0.23	0.38	0.31
K <sub>2</sub> O	<0.02	<0.02	0.10	0.02	<0.02	<0.02	<0.02	<0.02	<0.02	<0.02	<0.02	0.27	<0.02	<0.02	0.05	0.04
Total	90.60	92.92	80.47	64.78	76.03	79.54	81.06	90.53	76.73	81.66	85.47	94.34	97.95	77.53	87.48	87.75

All data in wt.%, by electron probe microanalysis; N = number of analyses; \* relic in scoriteaceous MM.

TABLE 2: COMPOSITION OF MINERAL PHASES IN CRYSTALLINE MICROMETEORITES.

N	M1		M6		M7		AM1		AM8		AM9		3M4									
	Ol	Px	Ol	Px	Ol	Plag**	Magnetite	Ol	Oxide	Oxide	Spinel**	Ol	Px	Sulfide*	Ol	Px	Oxide					
SiO <sub>2</sub>	41.6	51.9	56.7	<0.02	37.5	52.6	35.6	50.7	0.05	37.0	0.05	36.5	36.1	49.9	51.1	<0.02	39.6	57.8	53.4	8.70		
TiO <sub>2</sub>	0.08	0.08	<0.02	<0.02	0.04	0.13	<0.02	<0.02	0.05	0.03	0.02	0.02	0.02	0.74	0.72	<0.02	<0.02	0.09	0.08	0.27	0.19	
Al <sub>2</sub> O <sub>3</sub>	0.16	0.09	1.90	<0.02	0.02	0.24	0.07	32.1	0.35	0.03	0.03	2.40	1.82	26.5	0.03	1.30	2.71	<0.02	1.33	0.73	0.60	1.39
Cr <sub>2</sub> O <sub>3</sub>	0.19	0.06	<0.02	10.6	0.09	0.66	0.06	<0.02	3.40	<0.02	<0.02	5.50	4.60	9.60	0.12	0.85	0.40	0.10	0.50	0.61	0.72	2.63
FeO	2.11	12.8	2.90	67.0	17.9	11.3	25.8	<0.02	87.6	15.9	27.2	77.7	73.4	46.4	31.4	6.60	7.90	44.9	6.00	1.32	1.45	1.87
MnO	0.14	0.46	<0.02	<0.02	0.40	0.47	0.23	<0.02	0.05	0.43	0.21	0.10	0.18	<0.02	0.52	0.28	0.25	0.07	0.12	0.08	0.14	2.71
MgO	56.7	31.2	21.0	<0.02	43.8	31.3	37.3	<0.02	0.85	46.2	33.9	0.28	2.00	9.70	30.6	14.4	17.2	<0.02	52.2	38.0	37.8	25.7
NiO	0.05	<0.02	<0.02	5.80	0.03	<0.02	0.51	<0.02	0.14	0.27	<0.02	0.52	0.14	0.27	<0.02	12.3	0.05	<0.02	<0.02	<0.02	<0.02	2.38
CaO	0.21	0.29	17.6	<0.02	0.02	0.33	0.02	13.2	<0.02	0.04	0.03	0.08	0.19	1.50	0.58	21.4	17.3	<0.02	0.20	0.30	0.40	11.7
Na <sub>2</sub> O	<0.02	<0.02	<0.02	<0.02	<0.02	<0.02	<0.02	4.00	<0.02	<0.02	0.03	<0.02	0.04	1.28	0.44	<0.02	<0.02	<0.02	<0.02	<0.02	<0.02	0.05
Total	101.24	96.88	100.10	83.40	99.80	97.03	99.59	100.00	92.70	99.65	98.41	86.90	83.16	96.50	99.47	96.82	98.02	57.37	100.00	98.53	98.99	99.72

All data in wt%, determined by electron probe microanalysis; K<sub>2</sub>O in all phases  $\leq 0.02$  wt%; Px = pyroxene; Ol = olivine; Plag = plagioclase; \* elemental concentrations; \*\* EDS analyses (normalized to 100 wt%).

TABLE 3: CHEMICAL COMPOSITION OF PHASES FROM SCORIACEOUS MICROMETEORITES.

M2	M4		AM2		AM10		3M2		3M5		3M6B		4M5		4M12		
	Olivine	Olivine	Olivine	Fe-Oxide	Px	Px	Feldspar	Fe-Oxide	Px	Glass	Px	Olivine	Px	Fe-Oxide	Phyllo	Fe-Oxide	
N	4	3	2	4	1	1	1	1	1	2	1	2	1	3	2	2	
SiO <sub>2</sub>	35.8	40.4	35.5	39.8	7.20	58.8	57.6	51.0	65.8	1.69	57.7	55.8	31.0	57.9	40.7	56.9	40.9
TiO <sub>2</sub>	0.59	0.05	<0.02	<0.02	0.14	0.02	0.06	0.45	0.31	0.03	0.16	0.24	0.11	0.13	0.02	0.13	0.05
Al <sub>2</sub> O <sub>3</sub>	2.69	0.23	0.30	0.03	1.33	0.26	0.42	4.40	20.9	0.37	1.02	1.45	2.21	1.92	0.04	0.74	3.00
Cr <sub>2</sub> O <sub>3</sub>	0.36	0.33	0.40	0.37	0.22	0.82	0.82	2.10	0.06	0.16	0.46	0.88	0.90	0.60	0.42	0.42	0.64
FeO	26.1	1.29	27.7	0.79	72.8	1.98	2.95	2.00	1.28	74.6	1.51	3.30	38.2	0.53	1.32	1.31	78.9
MnO	0.30	0.15	0.21	0.26	0.16	0.20	0.24	2.68	1.21	0.03	0.10	0.55	0.37	0.13	0.11	0.12	16.0
MgO	30.7	58.2	36.3	57.4	7.20	39.1	38.1	19.5	1.04	0.43	37.1	33.4	24.1	37.7	55.8	39.1	17.5
NiO	1.07	<0.02	0.77	<0.02	1.05	0.08	0.18	<0.02	<0.02	1.40	<0.02	0.05	0.28	0.12	<0.02	0.22	1.33
CaO	0.75	0.32	0.24	0.13	0.11	0.26	0.34	16.8	4.60	0.22	0.51	2.21	0.40	1.88	0.21	0.50	0.08
Na <sub>2</sub> O	0.19	<0.02	<0.02	<0.02	0.03	<0.02	<0.02	0.14	7.80	0.30	<0.02	<0.02	0.33	<0.02	<0.02	<0.02	0.31
K <sub>2</sub> O	0.02	<0.02	<0.02	<0.02	<0.02	<0.02	<0.02	0.03	0.20	<0.02	<0.02	<0.02	<0.02	<0.02	<0.02	<0.02	<0.02
Total	98.57	100.97	101.42	98.78	90.24	101.52	100.71	99.10	103.20	79.23	98.56	97.88	97.90	100.91	98.62	99.44	84.65

All data in wt.%, by electron probe microanalysis; N = number of analyses; px = pyroxene, phyllo = phyllosilicate.

TABLE 4: CHEMICAL COMPOSITION OF BULK MICROMETEORITES.

Sample	Phyllosilicate MMs										Coarse-grained crystalline MMs										Scoriaceous MMs									
	M3	AM4	AM5	3M7	3M8	4M1	4M1	M1	M6	M7	AM1	AM8	M2	M4	M5	AM2	AM10	3M2	3M5	3M6	4M5	4M12	4M12	4M12	3	3	3			
N	7	9	9	3	5	8	8	7	5	6	19	3	5	5	4	8	5	8	5	4	8	8	3	3	3	3	3			
SiO <sub>2</sub>	23.9	25.3	25.7	25.2	35.2	29.0	53.2	53.2	50.1	32.3	39.7	34.7	21.2	24.7	26.5	29.5	21.1	27.5	30.4	22.1	26.7	37.8	27.60	37.8	27.60	37.8	27.60			
TiO <sub>2</sub>	0.08	0.09	0.09	0.09	0.08	0.09	0.13	0.13	0.16	0.03	0.08	0.01	0.08	0.05	0.07	0.06	0.07	0.07	0.07	0.07	0.07	0.06	0.05	0.04	0.05	0.04	0.05			
Al <sub>2</sub> O <sub>3</sub>	1.18	1.92	1.55	1.61	2.13	1.43	3.60	3.60	2.67	0.28	2.32	0.81	1.23	0.81	1.39	0.76	1.34	2.16	1.37	1.57	1.80	2.11	0.53	2.11	0.53	2.11	0.53			
Cr <sub>2</sub> O <sub>3</sub>	0.37	0.30	0.41	0.42	0.56	0.45	0.31	0.37	0.37	0.14	0.11	0.23	0.27	0.87	0.28	0.28	0.34	0.44	0.44	0.23	0.28	0.28	0.48	0.21	0.28	0.48	0.21			
FeO	20.2	33.8	26.5	24.0	22.4	28.2	11.9	12.9	12.9	26.8	13.0	27.6	24.9	29.7	24.1	13.5	25.9	29.6	28.5	34.3	32.2	19.8	35.90	19.8	35.90	19.8	35.90			
MnO	0.18	0.23	0.16	0.19	0.19	0.25	0.33	0.33	0.37	0.25	0.37	0.22	0.16	0.21	0.22	0.18	0.17	0.22	0.25	0.20	0.20	0.21	0.12	0.31	0.21	0.12	0.31			
MgO	15.7	13.7	12.3	13.5	15.2	14.5	23.7	23.7	29.4	29.0	34.4	30.2	14.0	21.2	15.0	30.1	13.3	12.6	14.9	13.2	14.2	22.6	17.00	22.6	17.00	22.6	17.00			
NiO	0.17	0.33	0.30	0.13	0.30	0.38	0.11	0.11	0.11	0.51	0.05	0.45	0.25	0.59	0.24	0.17	0.48	0.37	0.15	0.19	0.20	0.45	0.36	0.20	0.45	0.36				
CaO	0.26	0.31	1.21	0.72	1.41	0.88	2.58	2.58	1.43	0.53	2.01	0.68	0.28	0.39	0.86	0.67	0.64	1.22	1.78	0.54	1.06	0.23	0.79	0.23	0.79	0.23				
Na <sub>2</sub> O	0.16	0.18	0.35	0.17	0.24	0.21	1.18	0.59	0.03	0.58	0.08	0.08	0.12	0.02	0.14	0.07	0.14	0.69	0.24	0.12	0.16	0.30	0.05	0.16	0.30	0.05				
K <sub>2</sub> O	0.02	<0.02	0.01	0.07	0.18	0.04	0.04	0.01	<0.02	<0.02	0.07	<0.02	<0.02	<0.02	0.03	0.01	<0.02	0.14	0.07	0.03	0.04	<0.02	<0.02	<0.02	<0.02	<0.02				
SO <sub>2</sub>	0.69	1.05	1.84	0.63	1.43	2.03	0.39	0.39	0.10	0.45	0.05	0.05	0.22	0.29	0.51	0.85	0.30	0.88	0.90	0.35	0.57	1.65	0.67	0.57	1.65	0.67				
Total	62.90	77.21	70.42	66.73	79.38	77.46	97.46	98.21	90.33	92.73	95.03	62.71	78.83	69.33	76.15	63.79	75.87	79.07	72.90	77.47	84.02*	83.58**	84.02*	83.58**	84.02*	83.58**				

All data in wt.%, by electron probe microanalysis; N = number of analyses; \* includes 0.11 % Cl; \*\* includes 0.13 % Cl

TABLE 5: BULK CHEMICAL COMPOSITION OF MICROMETEORITES (DETERMINED BY NEUTRON ACTIVATION ANALYSIS).

Mass ( $\mu$ g)	Unmelted phyllosilicate MMs																	Anhydrous crystalline MMs																	Scleroaceous MMs																	Precision (rel.%)												
	AM4					AM5				AM7				AM1				AM8				AM9				AM4				AM2				AM10				AM12																										
	5	9	9	17	17	1	1	3	3	3	1	1	1	1	1	4	4	4	4	4	5	5	5	5	5	2	2	2	2	2	16	16	16	16	16	2	2	2	2	2	5	5	5	5	5	7	7	7	7	7	1.5		1.5	1.5	1.5	1.5	4	4	4	4	4	2	2	2
Na (wt%)	0.26	0.19	0.41	0.38	0.32	0.31	1.07	0.74	0.15	0.71	0.084	4.39	0.59	0.21	0.017	0.11	0.22	0.23	0.29	0.34	0.013	0.11	0.55	1-3																																								
K (wt%)	0.051	0.11	0.094	0.08	0.06	0.13	0.074	0.055	0.046	0.065	0.096	1.3	0.06	0.040	0.030	0.040	0.09	0.2	0.04	0.04	0.04	0.025	0.28	5-20																																								
Sc	10.2	4.78	7.93	11.4	11.3	5.63	17.8	10.4	5.68	17.1	5.72	19.6	43.9	8.75	18.4	7.02	19.1	5.28	4.09	8.66	2.72	4.03	13.3	1-4																																								
Cr	4380	2460	3890	5440	5510	2970	4270	3510	1410	1250	2570	4640	26400	3460	8190	3150	7050	3760	5080	1950	2453	8310	5-10																																									
Fe (wt%)	27.2	24.5	24.2	41.8	34.4	22.4	15.5	11.3	23.5	10.2	17.8	26.2	35.1	26.1	29.6	22.3	33.9	31.3	20.8	32.9	23.1	32.6	49.7	5-10																																								
Co	303	278	241	503	390	371	132	85	403	15.3	298	326	522	410	723	1225	683	619	222	482	310	419	846	1-5																																								
Ni	3890	3950	4560	8870	7560	3380	4050	3050	5810	610	3850	3410	10500	5510	12300	5840	6920	9440	4180	5060	3690	3060	10100	10-20																																								
Zn	145	128	95	345	255	350	90	70	75	72	107	67	107	19	25	1990	115	53	52	83	150	600	5-15																																									
Ga	8	9.5	3.5	0.3	0.5	6	1.5	5	3	1.7	4.2	12	12	6	10	9	5.7	19	1.6	0.2	0.6	5	2	10-20																																								
As	31	4.5	4.37	100	16	9.32	5.1	3.9	2.3	0.34	1.36	6.5	14	5.3	11.4	2.5	4.2	1.96	18	9.5	7.5	3.76	20.9	1-5																																								
Se	5.8	8.9	15.2	11	14.1	8.7	1.7	1.9	2.1	0.48	2.1	1.49	3	1.8	2.1	2.5	13.6	3.44	4.8	13.3	3.3	2.3	11.5	3-10																																								
Br	3.35	1.87	1.95	12.8	8.5	35.8	0.24	0.3	0.27	0.04	0.11	0.55	7.7	0.81	0.51	4.8	1.76	0.12	4.1	7.9	2.5	13.4	32.5	3-15																																								
Rb	6.2	2.5	5.4	<10	4.7	5.9	6.7	3.6	8.9	2.2	6.4	6.8	7.1	4.1	4.8	5.9	7.6	8.5	5.6	<10	3	2.1	<20	10-30																																								
Sr	<50	15	12	<250	<100	<100	<120	<100	15	8.1	<80	<140	<200	<70	<100	<100	<100	<150	<90	<200	<100	<120	<200	20-40																																								
Zr	6	7	20	<150	20	<100	<30	10	<50	10	12	<20	15	10	<40	<20	16	<20	30	<100	<50	<150	<500	25-40																																								
Ru	1.26	0.88	0.47	<1	0.8	2.1	0.52	0.68	0.35	0.46	0.53	0.55	3.8	0.42	0.49	0.45	0.83	0.77	0.39	2.9	0.6	<1.5	<2	5-15																																								
Sb	0.71	0.41	0.045	5.5	1.9	0.81	0.09	0.08	0.11	0.042	0.11	0.15	2.5	0.11	0.23	0.58	0.091	0.21	1.7	1.4	0.8	0.31	0.88	5-15																																								
Ba	<50	<30	4	<50	20	<25	<45	<30	<30	3.9	5.8	28	10	3.7	<20	6	<50	<40	15	<50	10	5	<80	25-40																																								
La	0.38	0.28	0.39	0.69	0.53	0.56	0.34	0.41	0.44	1.05	0.63	0.43	1.17	0.18	0.25	0.38	0.37	0.54	0.54	0.89	0.75	0.31	0.68	5-15																																								
Ce	n.d.	0.9	1	1.5	1.2	1.4	n.d.	1.2	n.d.	2.74	1.6	1.3	2.5	n.d.	n.d.	n.d.	1.2	1.5	1.3	2.2	1.6	0.7	1.8	15-25																																								
Nd	1	0.5	0.8	<3	0.8	<2	0.7	1.1	0.9	1.9	1.3	1.2	<2	0.4	0.35	0.65	<2	0.9	0.9	<4	0.9	<1	<4	25-40																																								
Sm	0.18	0.15	0.24	0.35	0.29	0.29	0.18	0.15	0.19	0.62	0.36	0.39	0.53	0.13	0.12	0.19	0.24	0.23	0.28	0.4	0.2	0.14	0.28	3-10																																								
Eu	0.064	0.064	0.1	0.16	0.11	0.11	0.075	0.068	0.064	0.15	0.11	0.14	0.12	0.063	0.046	0.065	0.075	0.1	0.095	0.13	0.11	0.08	0.1	3-10																																								
Gd	0.25	0.3	0.4	0.38	0.3	0.3	0.27	0.29	0.31	0.54	<0.5	0.55	0.45	0.24	0.2	0.28	<0.6	<0.4	0.35	0.4	0.3	0.3	<0.5	20-40																																								
Tb	0.056	0.068	0.09	<0.11	0.07	0.06	0.061	0.058	0.064	0.11	0.07	0.12	0.085	0.043	0.041	0.044	<0.1	0.07	0.11	0.1	0.057	0.06	<0.1	10-35																																								
Tm	<0.5	0.06	0.06	<0.1	<0.05	<0.3	<0.1	0.04	<0.1	0.08	<0.1	0.07	<0.1	<0.1	<0.3	<0.1	<0.2	<0.1	0.05	<0.1	<0.08	<0.01	<0.1	25-40																																								
Yb	0.23	0.31	0.37	0.4	0.19	0.38	0.24	0.26	0.31	0.49	0.29	0.43	0.75	0.23	0.24	0.28	0.35	0.38	0.36	0.39	0.4	0.31	0.3	10-30																																								
Lu	0.042	0.042	0.056	0.055	0.03	0.05	0.034	0.039	0.042	0.079	0.028	0.062	0.045	0.036	0.034	0.045	0.053	0.052	0.051	0.053	0.056	0.045	0.03	5-20																																								
Hf	0.47	0.15	0.08	<0.5	0.3	<0.5	0.22	0.24	0.31	0.34	0.12	0.34	1.4	0.28	0.41	0.28	0.23	0.28	0.38	0.42	0.35	<0.5	<2	10-25																																								
Ta	0.11	0.044	0.034	0.1	0.4	<0.05	0.07	0.04	0.28	0.021	0.055	0.04	1.3	0.12	0.11	0.11	0.45	0.11	0.41	0.2	0.08	<0.08	<0.5	20-35																																								
Os (ppb)	625	360	380	1050	670	500	320	100	260	270	124	230	1100	310	320	445	800	503	450	600	490	462	975	10-25																																								
Ir (ppb)	712	433	446	895	851	410	187	142	297	265	144	222	2050	470	789	344	958	536	480	689	305	452	1110	2-6																																								
Au (ppb)	1260	320	263	710	680	278	180	41	125	50	27	87	209	139	218	338	617	142	195	403	62	94	235	1-5																																								
Th	<0.21	0.069	0.11	<0.3	<0.3	<0.3	<0.21	0.031	<0.1	0.024	0.035	0.11	0.35	0.025	<0.38	<0.2	0.04	0.18	0.62	0.19	0.23	<0.2	<1	20-35																																								
U	<0.4	0.09	0.015	<1	0.4	<0.1	0.11	0.05	<0.12	0.052	0.11	0.09	0.38	0.12	<0.14	<0.2	0.21	0.08	0.41	<0.5	0.09	<0.1	<0.3	25-40																																								

All data in ppm, except as indicated; Analyses have various levels of precision (indicated in the column on the right) as a function of sample weight, counting parameters, interferences, and other factors.

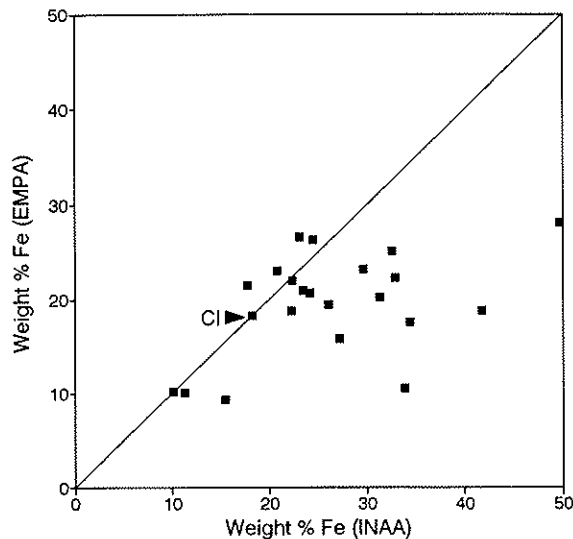


FIG. 4. Comparison of bulk Fe contents of micrometeorites as determined by EMPA vs. those determined by INAA. Bulk data derived from analyzing the total micrometeorite deviate considerably from those derived by EMPA by analyzing the interior which in most cases contains much less Fe than the bulk.

are typical of low-Ca pyroxenes from carbonaceous chondrites (e.g., HOINKES and KURAT, 1975; NELEN et al., 1975; KURAT, 1975; BRANDSTÄTTER et al., 1992). Only two FeO-rich (Fs 17–23) low-Ca pyroxenes were encountered. They have different contents of minor elements but both have low FeO/MnO ratios indicating a carbonaceous chondrite heritage. A pigeonite grain in 3M6 has a chemical composition compatible with a carbonaceous chondrite heritage, except for its high NiO content. However, considering the small size of this grain, the high Ni reading could be due to excitation of a nearby Ni-rich grain. One clinopyroxene grain with an extremely low FeO/MnO ratio (0.75) is present in scoriaceous MM 3M2 (Table 3). Such low Fe/Mn ratios are known only from unequilibrated ordinary chondrites, carbonaceous chondrites, and SIDPs (NELEN et al., 1975; HUTCHISON, 1987; KLÖCK et al., 1989). Thus, pyroxene compositions indicate a general relationship to CM and CR chondrites. However, the lack of very low-Fe pyroxenes, and the high abundance of pyroxene, is not typical of known carbonaceous chondrites.

Plagioclase is occasionally present in crystalline and scoriaceous MMs and ranges in composition from An 12 to An 25, which is not typical of carbonaceous chondrites but still within the range of plagioclase compositions in these meteorites. The association of plagioclase An 12 with a Fa-rich olivine in MM AM9 could indicate a relationship to a highly oxidized ordinary chondrite type.

Magnetite is occasionally present in a variety of morphologies similar to those known from CM and CI chondrites (KURAT et al., 1992b). In one case (M7) it is present in a CK chondrite association. More common are iron oxides of nonmagnetite composition, some with affinities towards tochilinite, the dominating iron oxide of CM chondrites (e.g.,

ZOLENSKY, 1987). In addition, ferrobrucite (amakinite) is occasionally present, which is also typical of CM chondrites (ZOLENSKY and MCSWEEN, 1988). However, micrometeorites commonly contain a peculiar iron oxide phase rich in C, O, P, S, and other elements, which has been dubbed COPS (PERREAU et al., 1992; ENGRAND et al., 1993). This phase occurs within voids of MMs or at their surface, on top of the magnetite crust. It is also present at the surface of some cosmic spherules. Therefore, it must be of a post-melting origin—consequently of terrestrial origin. Work aimed at characterizing the COPS phase in proper detail is under way.

Metal phases are rare and present either as droplet inclusions in olivines (M1) or in the glassy matrix (4M5). They are low in Ni, which is typical of primary metal as it is present in Renazzo (e.g., NELEN et al., 1975; LEE et al., 1992) and other chondrites (PERNICKA et al., 1985, 1989). Sulfides are rare in micrometeorites but, if present, are either low-Ni pyrrhotite or high-Ni pentlandite, both of which are typical of carbonaceous chondrites.

Several phases that are abundant in CM chondrites are not present in micrometeorites. No magnesium sulfates or (Mg, Na) sulfates and no calcium sulfates have been encountered, which are typical of hydrated carbonaceous chondrites (e.g., BOSTROEM and FREDRIKSSON, 1966; FREDRIKSSON et al., 1980; ZOLENSKY and MCSWEEN, 1988). Also, none of the carbonates common in CM chondrites (calcite, magnesite, dolomite, ankerite, and kutnahorite; e.g., BRANDSTÄTTER et al., 1987; FREDRIKSSON and KERRIDGE, 1988; JOHNSON and PRINZ, 1993) have been encountered. Some of the elemental depletions in MMs appear to be directly related to the absence of these minerals (see below).

In summary, the mineralogical and mineral chemical composition of micrometeorites has a clear relationship to that of hydrated carbonaceous chondrites, particularly the CM and CR chondrites. The match, however, is not perfect and, thus, argues against a direct derivation of MMs from CM and CR-like rocks. The differences between MMs and CM and CR chondrites are considerable and they appear partly to be due to terrestrial alterations of the MMs, but some of them must be primordial. Differences most probably due to terrestrial alteration are the lack of sulfates and carbonates, and the presence of COPS (plus the magnetite crusts, see below). Sulfates and carbonates are soluble in low pH water. Considering the fact that all precipitation from the terrestrial atmosphere is acid, this criterion is easily met. Thus, sulfates and carbonates could have been lost by dissolution in the Antarctic ice and in the melt ice water during dust recovery. A variety of elements could have been dissolved with the sulfates and carbonates which are now missing in MMs (Ca, Ni, S, Na, and also Mg and Mn; e.g., KURAT et al., 1992a,b; PRESPEL et al., 1993; see discussion below). COPS is a phase not found in meteorites but is found in meteorite fusion crusts (ENGRAND et al., 1993) and is of terrestrial origin. It probably forms by condensation in the upper atmosphere from vapors formed by evaporation of micro- and micrometeoroids. The terrestrial E-layer has been shown to be particularly rich in elements evaporated from meteoroids (e.g., STEINWEG et al., 1992).

Apart from terrestrial alteration, some differences in min-

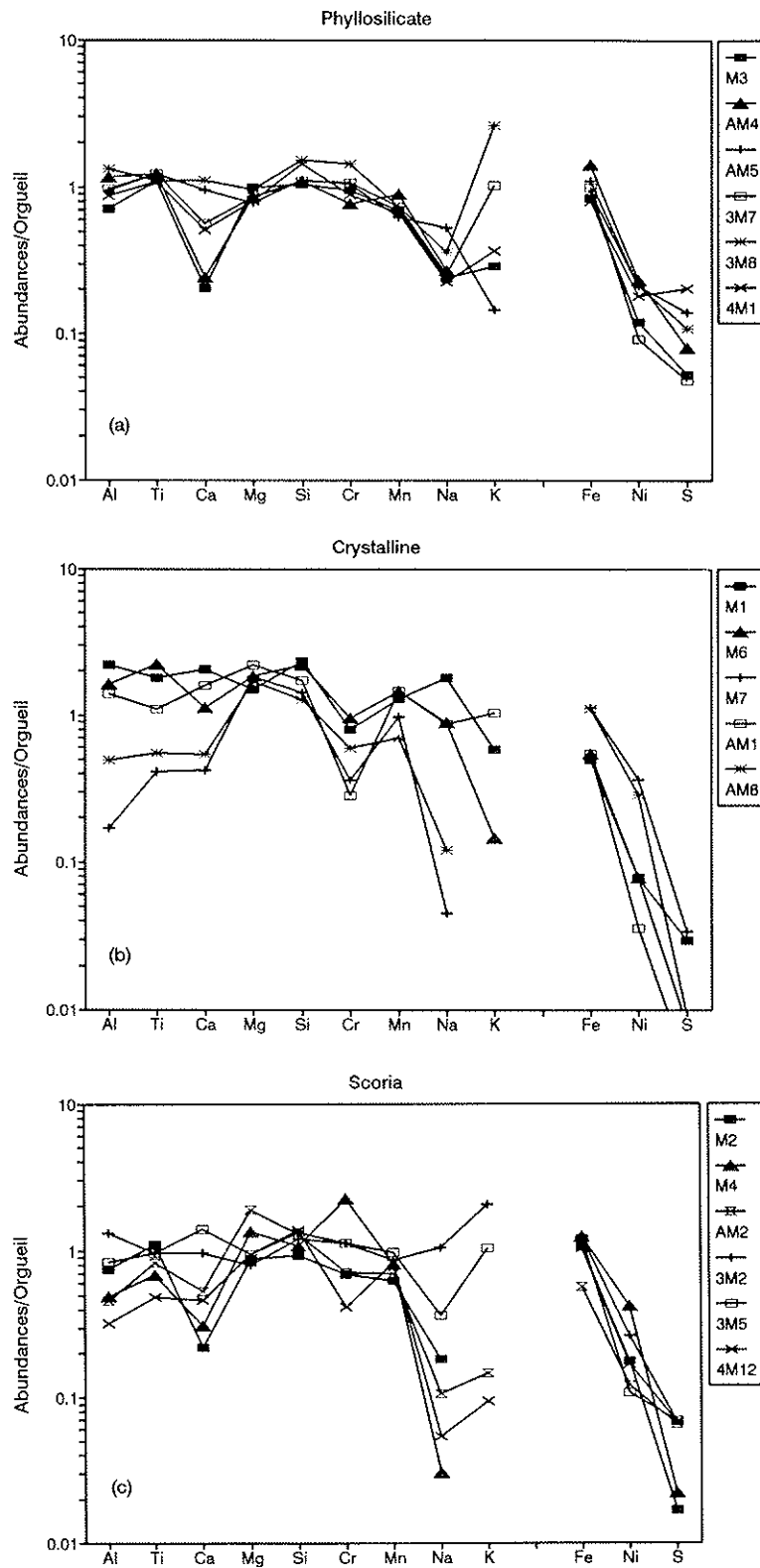


FIG. 5. Abundances of major and minor elements in micrometeorites (EMPA data, Table 4) normalized to CI abundances (PALME et al., 1981). Elements are separated into lithophile (left) and siderophile (right) elements and arranged in order of increasing cosmochemical volatility from left to right (WASSON, 1985). (a) Phyllosilicate MMs; (b) Crystalline MMs; (c) Scoriaceous MMs.

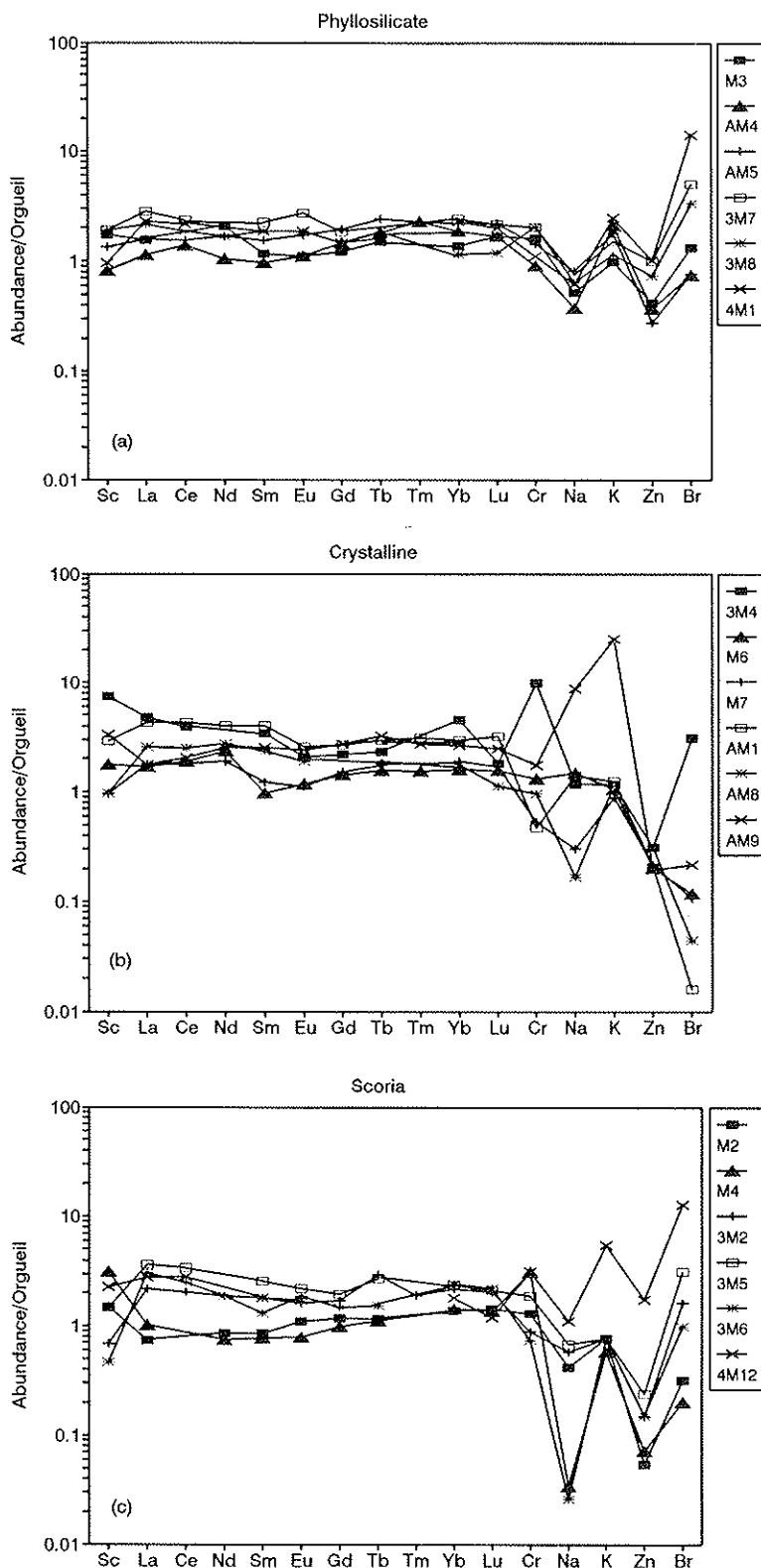


FIG. 6. Abundances of selected lithophile elements in micrometeorites normalized to CI abundances (PALME et al., 1981). Elements are arranged in order of increasing cosmochemical volatility from left to right (WASSON, 1985), with the exception of the rare earth elements, which are arranged according to increasing atomic number. (a) Phyllosilicate MMs; (b) Crystalline MMs; (c) Scoriaceous MMs.

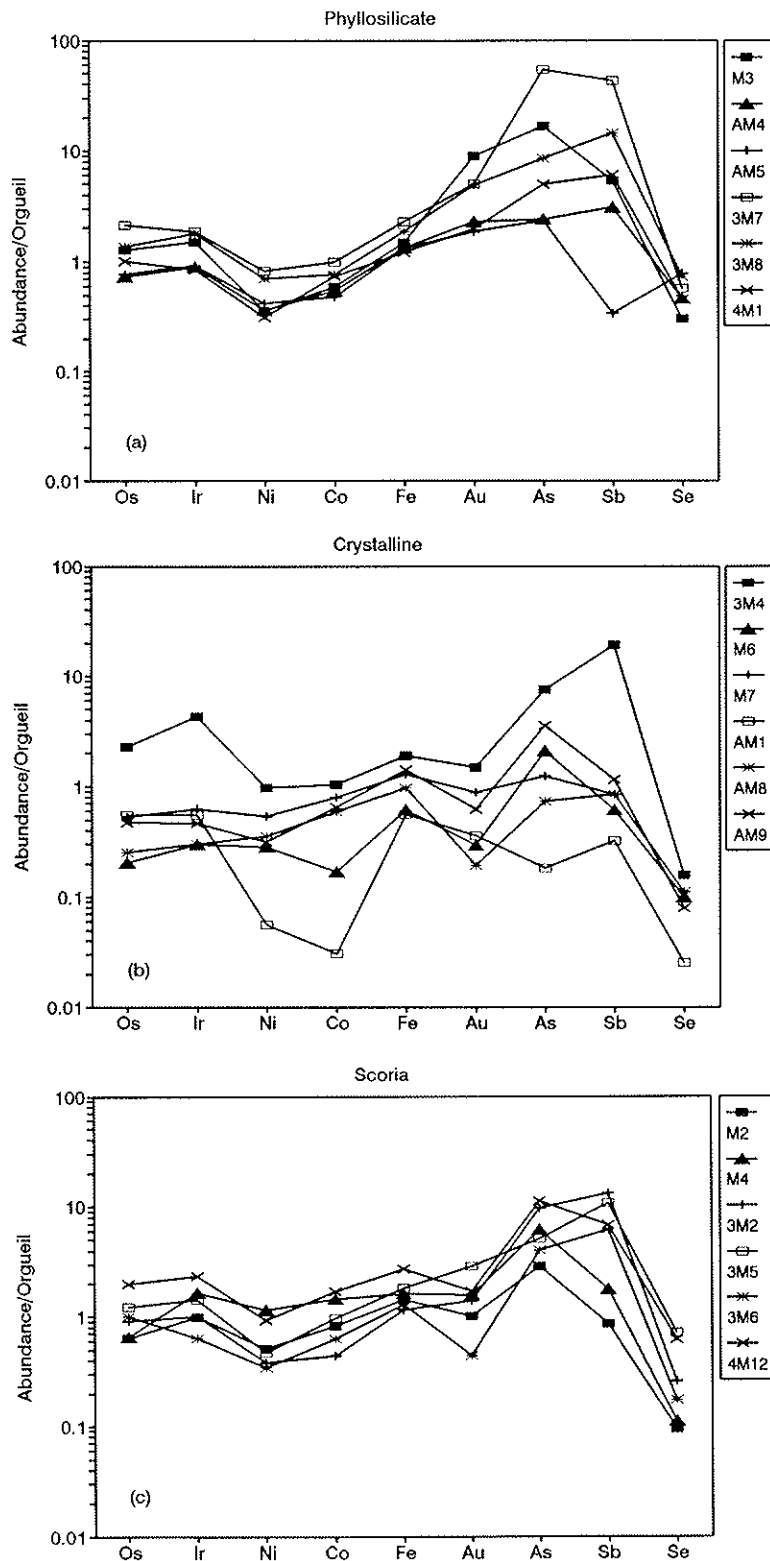


FIG. 7. Abundances of selected siderophile elements in micrometeorites normalized to CI abundances (PALME et al., 1981). Elements are arranged in order of increasing cosmochemical volatility from left to right (WASSON, 1985). (a) Phyllosilicate MMs; (b) Crystalline MMs; (c) Scoriaceous MMs.



eralogy and mineral chemistry between micrometeorites and CM and CR chondrites remain. They point towards a somewhat different nebular processing of matter finally ending up in either carbonaceous chondrites or interplanetary dust. The processes which were involved were similar in both cases but they appear to have been operative to differing degrees in creating these different sample populations.

The low ol/px ratio for MMs indicates a high production rate of pyroxene and thus a more efficient conversion of olivine into pyroxene (e.g., KURAT, 1988). The lack of very Fe-poor, refractory element-enriched olivines in MMs points into the same direction because apparently none of the olivines escaped processing (annealing and crystal-vapor exchange of elements).

Most of the accessory phases in MMs, such as Mg,Al-spinel, Fe,Al-spinel, perovskite, ilmenite, fassaite, hibonite, and others, are similar to those found in carbonaceous chondrites. Because the MM sample set is much smaller than that of the CM and CR chondrites, it cannot be expected that all of the phases present in chondrites have already been detected in MMs.

None of the MMs investigated so far has an unambiguous mineral chemical relationship to either ordinary chondrites or achondrites with the possible exception of aubrites (compare 3M4, Fig. 2a,b). However, all silicates of the coarse-grained crystalline MMs are too rich in FeO to qualify for any of the known enstatite meteorites. Such high FeO contents (and correspondingly high MnO contents) indicate extensive crystal-vapor exchange reactions. Thus, the coarse-grained crystalline MMs also appear to have experienced processing similar to that of the isolated mineral grains of the phyllosilicate MMs. This view is strongly supported by the presence of tochilinite-like oxides in crystalline MMs. The voids which are abundantly present in crystalline MMs are likely to be the result of dissolution of metals and common sulfides rather than the exotic sulfides of the enstatite meteorite clan. Therefore, it appears likely that most of the micrometeorites belong to just one reservoir. The only likely exceptions are the olivine of M2 (Table 3) and the CK chondrite assemblage of M7 (Table 2).

Comparison of the mineralogy and mineral chemistry of the micrometeorites with those of the SIDPs reveals some similarities but also some important differences. The major silicates of the SIDPs are similar to those in MMs—phyllosilicates, olivine, and pyroxene. In both cases the phyllosilicates form a low-porosity matrix which eventually encloses grains of anhydrous silicates and other minerals (e.g., BROWNLEE et al., 1989). The nature of the phyllosilicates, however, is different: SIDPs are dominated by smectite (e.g., BRADLEY, 1988; ZOLENSKY and LINDSTROM, 1991), while MMs (and CM and CR chondrites) are dominated by serpentine (e.g., BUNCH and CHANG, 1980; BARBER, 1985; ZOLENSKY and MCSWEEN, 1988; WEISBERG et al., 1993; ZOLENSKY et al., 1993). Only a few SIDPs contain serpentine (THOMAS et al., 1990). Furthermore, cronstedtite is rare in SIDPs (RIETMEIJER, 1992), but tochilinite is common (BRADLEY and BROWNLEE, 1991; KLÖCK et al., 1992b).

In addition, the anhydrous silicates in SIDPs and MMs appear to be fundamentally different. In SIDPs olivine and pyroxene form fine-grained porous aggregates (e.g., BROWN-

LEE, 1987), or are finely dispersed as small grains within the phyllosilicate matrix (e.g., BRADLEY et al., 1988). Such aggregates have not been found in the MMs. Another apparent difference between the SIDPs and MMs is the variety of weird phases which have been reported from SIDPs: Sn-rich grains, Al-, Si-, Ti-, Bi-oxides, SiC, phosphides, Fe-Cr sulfide, and BaSO<sub>4</sub> (MACKINNON and RIETMEIJER, 1984; RIETMEIJER and MACKINNON, 1984a,b, 1985, 1990; RIETMEIJER, 1985; RIETMEIJER and MCKAY, 1986). Some, but not all, of these phases also occur in our MM sample, but they are probably not indigenous.

### Bulk Chemistry

Considering the small masses of the micrometeorites, their bulk chemical compositions are surprisingly similar to one another with the least spread exhibited by the phyllosilicate MMs. Abundances of refractory, moderately volatile, and volatile lithophile elements in these MMs (Figs. 5, 6) match, in general, the abundance patterns of CM chondrites, but not that of CR chondrites (WASSON and KALLEMEYN, 1988). In many cases Ca, Na, and Zn are depleted in MMs as compared to CM chondrites, and K and Br are enriched. The siderophile and chalcophile elements (Figs. 5, 7) are fractionated, except for the refractory elements Os and Ir and, occasionally, Se. Nickel and Co are depleted in MMs as compared to CI (and CM) chondrites, and Fe, Au, As, and Sb are enriched. Some of the elemental depletions and enrichments observed for MMs have also been found in SIDPs. Originally most of these positive and negative deviations from the chondritic composition, especially those encountered in SIDPs, were interpreted as being primary (e.g., VANDER STAP et al., 1986; FLYNN and SUTTON, 1989, 1990, 1991, 1992a; SUTTON and FLYNN, 1988, 1989; SCHRAMM et al., 1989; THOMAS et al., 1992). Primary enrichments were assumed, especially for the volatile elements S, Mn, Cu, Zn, Ga, Fe, Se, and Br, but also for C, Y, Nb, and Mo. As a consequence, SIDPs were considered to be samples of solar nebula matter enriched in volatile elements, perhaps transitional between CI chondrites and cometary matter (e.g., FLYNN and SUTTON, 1992a). However, JESSBERGER et al. (1992) showed that most of these elemental enrichments in SIDPs are the result of contamination from the terrestrial atmosphere, especially from the E-layer (e.g., HUNTEN et al., 1980; STEINWEG et al., 1992; KANE and GARDENER, 1993). We believe that the same holds true for the micrometeorites. Consequently, the enrichments of MMs in K, Fe, As, Br, Rb, Sb, and Au, as compared to CI chondrites, are likely to be of terrestrial atmospheric origin (e.g., KURAT et al., 1992a,b, 1993). The reconcondensation of Fe is evident from magnetite envelopes and Fe metasomatism (Figs. 1–3, 8). Also, comparison of the bulk compositions of particles to their interior compositions shows surface-related elemental enrichments (Fig. 4). As the magnetites coating the MMs are also rich in other elements, they also add to the bulk composition. However, the abundances of these elements in magnetite are commonly too low to show up in the bulk analyses (with the exception of Ni).

A secondary enrichment of highly volatile elements in MMs is also suggested by the fact that, especially in the scori-

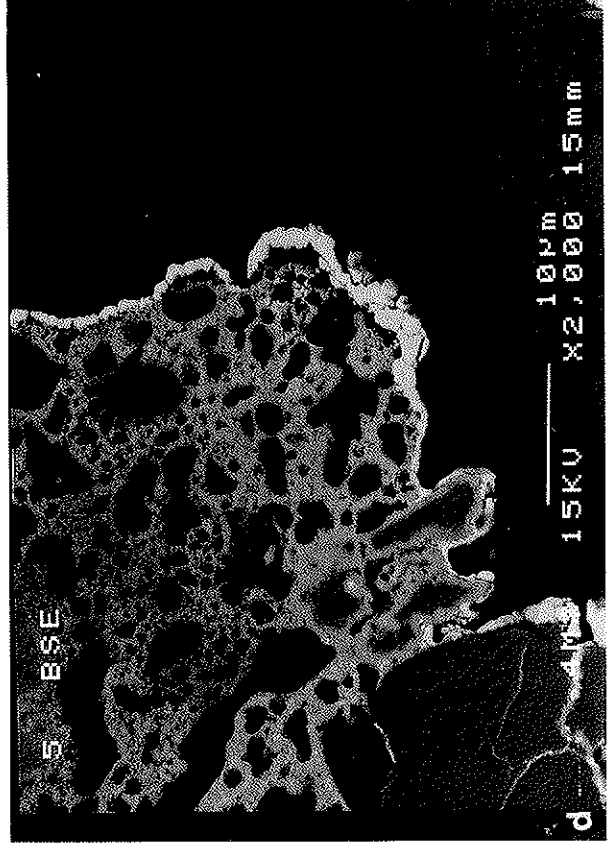


TABLE 6: COMPOSITION OF MAGNETITE COATINGS ON MICROMETEORITES.

Object	M2	M3	M4		M4		AM9	3M8
			large plates		small octahedra			
SiO <sub>2</sub>	7.4	5.2	3.1	6.1	2	2.2	7.6	3
TiO <sub>2</sub>	<0.1	<0.1	0.3	0.2	0.8	0.7	0.3	<0.1
Al <sub>2</sub> O <sub>3</sub>	5.5	10	2.9	5	8.7	8.2	3.4	2.1
Cr <sub>2</sub> O <sub>3</sub>	<0.1	<0.1	0.8	0.5	0.3	0.4	<0.1	0.5
FeO	76	69	78	75	80.4	80.5	82.4	86.1
MnO	<0.1	<0.1	1.1	1.1	0.9	0.9	0.5	<0.1
MgO	7.2	12.1	7.4	6.2	4.1	4.1	6.6	6.3
NiO	3.9	3.6	8.7	7.7	4.8	5	<0.1	2.5

All analyses by EDS and normalized to 100 wt%; all Fe as FeO.

aceous MMs, Na is depleted, but K and other elements which are much more volatile than Na are not depleted. Because the Na contents decrease systematically from unmelted to scoriaceous MMs to cosmic spherules, this decrease must be due to volatilization during atmospheric heating. For reasons unknown to us, Na does not tend to recondense onto the cooled particles at the lower base of the E-layer but stays in the atmosphere for a prolonged time in the "sodium-layer" (e.g., GADSDEN, 1968; FERGUSON, 1978; KANE and GARDENER, 1993) and from there it is removed by a mechanism which must be different from that for most other elements.

Only one element commonly enriched in MMs and SIDPs over CI levels could represent a primary enrichment: carbon. There is a common phase rich in C which is of terrestrial atmospheric origin (COPS) but PERREAU et al. (1993) estimate its abundance at only 5% of all C-rich phases encountered. This indicates that either the excess C in MMs is primary, or that secondary (atmospheric) C-rich phases do not always form COPS. This problem needs further studies, which are under way.

The elemental depletions commonly encountered in MMs and SIDPs have been a mystery for some time. The occasionally low abundances of Ca were considered to be a primary feature of SIDPs (e.g., FLYNN and SUTTON, 1989, 1992a; SCHRAMM et al., 1989) and MMs (MAURETTE et al., 1991, 1992c). However, Ca depletions, together with depletions in Na, S, Ni, and sometimes also Mg and Mn, are typical of the silicate portion of CM matrices (e.g., SCOTT et al., 1988; NAZAROV et al., 1993, Fig. 9). This is due to the fact that these elements preferentially enter nonsilicate phases such as carbonates (Ca, Mg, Mn; e.g., BRANDSTÄTTER et al., 1987; FREDRIKSSON and KERRIDGE, 1988; JOHNSON and PRINZ, 1993), sulfates (FREDRIKSSON et al., 1980; PRESPEER et al., 1993), sulfides, and chlorides. All of these phases form either large poikiloblasts or fill cracks and therefore are usually excluded from matrix bulk analysis. Nevertheless, depletions in S, Ni, and Zn in SIDPs have been interpreted as being due to evaporative loss during atmospheric entry heating (e.g.,

FLYNN and SUTTON, 1989, 1992b,c; THOMAS et al., 1992). We consider such a mechanism unlikely and have, for similar depletions observed in MMs, proposed losses of soluble phases (carbonates, sulfates, halogenides) from MMs by leaching in the atmosphere, the ice, and the melt ice water (e.g., KURAT et al., 1992a,b, 1993; PRESPEER et al., 1993).

### Magnetite Envelopes

Magnetite envelopes have been noted on SIDPs (e.g., THOMAS et al., 1992; KLÖCK et al., 1992b) and were related to the atmospheric heating event. Such envelopes are commonly present around both unmelted and scoriaceous MMs (e.g., MAURETTE et al., 1991; KURAT et al., 1992a). Their morphologies (Figs. 3, 8) and chemical compositions (Table 6, rich in Si, Al, Mg, Ni) strongly suggest an origin by condensation rather than by simple "oxidation of extraterrestrial objects in the atmosphere," as suggested by ROBIN et al. (1992). Condensation onto cooling extraterrestrial particles apparently provides an important sink for the large amounts of Fe vapor formed by evaporation of meteoroids in the high atmosphere (e.g., STEINWEG et al., 1992). The widespread occurrence of such magnetite envelopes makes it clear that this is a common process in the atmosphere. Consequently, such magnetites (the first refractory aerominerals described!) must have formed throughout geological time, provided the terrestrial atmosphere was oxidizing to a degree comparable to today's air. Magnetites should, therefore, be present in all Phanerozoic sediments where they are indeed common (e.g., MUTCH, 1966; BROWNLEE, 1985; IWAHASHI et al., 1991). However, isolated magnetites of platy and octahedral morphologies similar to those covering micrometeorites have so far been found only at a few locations and mainly in the K/T boundary layer (KYTE and SMIT, 1985; ROBIN et al., 1992). Their presence has been taken as a support of the impact-induced mass extinction of biota at the K/T boundary. However, the omnipresence of such magnetites among micrometeorites considerably weakens that argument.

### CONCLUSIONS

Micrometeorites represent the major fraction of the matter accreting to the Earth today. Particles with sizes between 100 and 400  $\mu\text{m}$  consist mostly of matter similar in mineralogy and chemical composition to the constituents of CM carbonaceous chondrites, with some features also being comparable to CR and CI chondrites. However, their proportion in the total interplanetary dust sample is not known mainly because other possible components with nonchondritic properties have not yet been identified. Presumably such a nonchondritic component is present in the Antarctic dust

FIG. 8. Scanning electron microphotographs of magnetite envelopes of micrometeorites. (a) Large platy magnetite covered by smaller octahedral magnetite at the surface of scoriaceous MM M4. SE image. (b) Magnetites lining the walls of an open vesicle at the surface of scoriaceous MM M4. SE image. (c) High magnification of octahedral magnetite with a generation of very small magnetites at the surface of scoriaceous MM M4. SE image. (d) Polished section of scoriaceous MM 4M5, showing a highly vesicular melt with relic olivines and pyroxenes (dark grey). The surface is covered by granular magnetite (white). Note that the magnetite is present everywhere, independent of the substrate. BSE image.

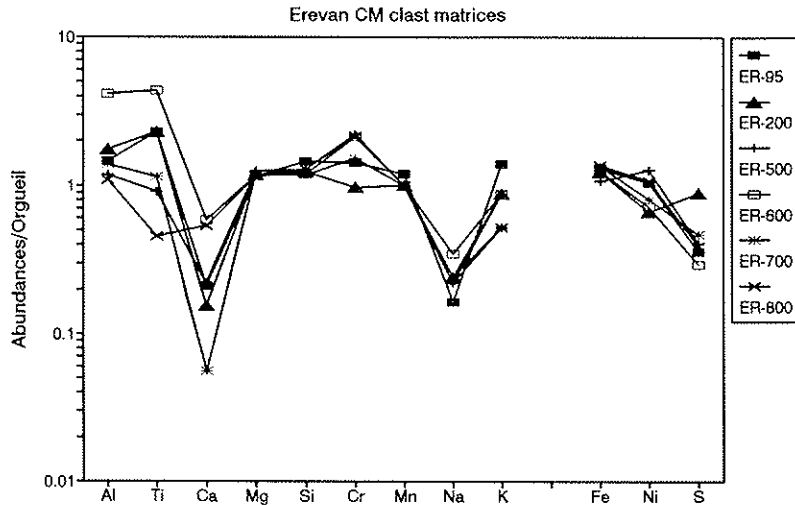


FIG. 9. Abundances of major and minor elements in matrices of CM chondrite inclusions in the Erevan howardite (NAZAROV et al., 1993), normalized to CI abundances (PALME et al., 1981). Elements are shown in the same order as in Fig. 5 to facilitate comparison.

collection and its identification will require a different approach.

The chondritic component of the dust in the 100–400  $\mu\text{m}$  size range consists of phyllosilicate-dominated and anhydrous mineral-dominated unmelted micrometeorites and partly melted scoriaceous micrometeorites. The latter clearly dominate this size range of Antarctic dust but the proportion of unmelted phyllosilicate MMs is surprisingly high. Of the total population of phyllosilicate-rich meteoroids (phyllosilicate plus scoriaceous MMs, 16 particles), 38% (by number) survived atmospheric entry without melting. This is an astonishingly large proportion even considering that most of these unmelted MMs were thermally metamorphosed by the frictional heat accompanying atmospheric entry. The survival of pristine phyllosilicates is less common.

The primary mineral association of chondritic micrometeorites has probably only a single source. The combination of phyllosilicates with anhydrous coarse-grained minerals (olivine and pyroxene) points towards a source similar to that of CM carbonaceous chondrites, a rare chondrite type ( $\sim 2\%$  of all chondrite falls, DODD, 1981). The quantitative fit is not so good if the mineral proportions and mineral chemical compositions are considered: micrometeorites have ol/px ratios of  $\sim 1$  and olivine and pyroxene compositions which lack the Fe-poor, refractory element-rich portion characteristic of CM chondrites. Thus, micrometeorites consist of components, which have experienced similar processing as those of CM chondrites, but this processing apparently was of higher intensity, e.g., solid-gas exchange processes in the solar nebula proceeded further in the components of MMs than that of CM chondrites (compare KURAT, 1988). The differences in mineral abundances and composition are probably pristine. Another obvious difference in mineralogy is the lack of carbonates and sulfates in MMs, minerals which are major constituents of CM chondrites. This feature is very likely of terrestrial origin as sulfates and carbonates were probably lost by dissolution in the ice or melt-ice water.

The dissolution of soluble phases had some effects on the

bulk chemical compositions of MMs. Depletions of MMs in Na, Mg, Ca, Mn, Ni, Co, and S are probably the result of terrestrial leaching of sulfates and carbonates. The chemical compositions of unmelted MMs are otherwise similar to that of CM chondrites, including the abundances of diagnostic volatile trace elements such as Zn and Br. However, some volatile and other elements are commonly overabundant in MMs as compared to CM and CI chondrites. Such enrichments in K, Br, Rb, As, Sb, Au, and also Fe are also likely of terrestrial origin. These elements probably recondensed from meteoric vapors in the atmospheric E-layer. In particular, the condensation (and metasomatic infiltration) of Fe (diluted by some elements like Si, Al, Ni) is documented in Fe-rich rinds and magnetite envelopes around most micrometeorites. Although some enrichment of MMs in C is visible in the deposits of COPS in voids and at the surface, the source of the overall large enrichment in C is not known. The C-rich nature of MMs (and SIDPs) could therefore be a primary feature. Studies of MMs should also be of interest to terrestrial atmosphere research because they carry information on their interaction with the terrestrial atmosphere at high and low temperature.

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## REFERENCES

- AKAI J. (1988) Incompletely transformed serpentine-type phyllosilicates in the matrix of Antarctic CM chondrites. *Geochim. Cosmochim. Acta* **52**, 1593–1599.

- ALEXANDER C. M. O'D, MAURETTE M., SWAN P., and WALKER R. M. (1992) Studies of Antarctic micrometeorites. *Lunar Planet. Sci.* **23**, 7–8.
- BARBER D. J. (1985) Phyllosilicates and other layer-structured materials in stony meteorites. *Clay Minerals* **20**, 415–454.
- BECKERLING W., BISCHOFF A., and KLÖCK W. (1992) Mineralogy and chemistry of micrometeorites from Greenland and Antarctica. *Meteoritics* **27**, 200–201.
- BOSTROEM K. and FREDRIKSSON K. (1966) Surface conditions of the Orgueil meteorite parent body as indicated by mineral associations. *Smithson. Misc. Coll.* **151/3**, 1–39.
- BRADLEY J. P. (1988) Analysis of chondritic interplanetary dust thin-sections. *Geochim. Cosmochim. Acta* **52**, 889–900.
- BRADLEY J. P. and BROWNLEE D. E. (1991) An interplanetary dust particle linked directly to type CM meteorites and an asteroidal origin. *Science* **251**, 549–552.
- BRADLEY J. P., SANDFORD S. A., and WALKER R. M. (1988) Interplanetary dust particles. In *Meteorites and the Early Solar System* (ed. J. F. KERRIDGE and M. S. MATTHEWS), pp. 861–895. Univ. Arizona Press.
- BRANDSTÄTTER F., KURAT G., and GRAHAM A. L. (1987) Primitive carbonates in Y82042 (C2). *Meteoritics* **22**, 336–337.
- BRANDSTÄTTER F., KURAT G., and MAURETTE M. (1991) Pauschal- und Mineralchemismen antarktischer Mikrometeorite. *Beih. Eur. J. Mineral.* **3/1**, 40.
- BRANDSTÄTTER F., KURAT G., and IVANOV A. V. (1992) Isolated minerals in Kaidun II (CI). *Meteoritics* **27**, 206.
- BROWNING L., ZOLENSKY M., and BARRET R. (1991) Serpentine and modal compositions of CM chondrites. *Lunar Planet. Sci.* **XXII**, 145–146.
- BROWNLEE D. E. (1981) Interplanetary dust—its physical nature and entry into the atmosphere of terrestrial planets. In *Comets and the Origin of Life* (ed. C. PONNAMPERUMA), pp. 63–70. Reidel.
- BROWNLEE D. E. (1985) Cosmic dust: collection and research. *Ann. Rev. Earth Planet. Sci.* **13**, 147–173.
- BROWNLEE D. E. (1987) Morphological, chemical and mineralogical studies of cosmic dust. *Phil. Trans. Roy. Soc. London* **A323**, 305–311.
- BROWNLEE D. E., SCHRAMM L. S., WHELOCK M. W., and MAURETTE M. (1989) Large mineral grains in interplanetary dust. *Lunar Planet. Sci.* **XX**, 121–122.
- BUNCH T. E. and CHANG S. (1980) Carbonaceous chondrites—II. Carbonaceous chondrite phyllosilicates and light element geochemistry as indicators of parent body processes and surface conditions. *Geochim. Cosmochim. Acta* **44**, 1543–1577.
- CHRISTOPHE MICHEL-LEVY M. and BOUROT-DENISE M. (1992) Mineral compositions in Antarctic and Greenland micrometeorites. *Meteoritics* **27**, 73–80.
- DODD R. T. (1981) *Meteorites. A Petrologic-Chemical Synthesis*. Cambridge Univ. Press.
- ENGRAND C., MAURETTE M., KURAT G., BRANDSTÄTTER F., and PERREAU M. (1993) A new carbon-rich phase (“COPS”) in Antarctic micrometeorites. *Lunar Planet. Sci.* **XXIV**, 441–442.
- ESSER B. K. and TUREKIAN K. K. (1988) Accretion rate of extraterrestrial particles determined from osmium isotope systematics of Pacific pelagic clay and manganese nodules. *Geochim. Cosmochim. Acta* **52**, 1383–1388.
- FERGUSON E. E. (1978) Sodium hydroxide ions in the stratosphere. *Geophys. Res. Lett.* **5**, 1035–1038.
- FLYNN G. J. and SUTTON S. R. (1989) Minor and trace element abundances in eight “chondritic” stratospheric particles: Evidence for Ni depletions. *Meteoritics* **24**, 267.
- FLYNN G. J. and SUTTON S. R. (1990) Synchrotron X-ray fluorescence analyses of stratospheric cosmic dust: New results for chondritic and low-nickel particles. *Proc. 20th Lunar Planet. Sci. Conf.*, 335–342.
- FLYNN G. J. and SUTTON S. R. (1991) Average minor and trace element contents in seventeen “chondritic” IDPs suggest a volatile enrichment. *Meteoritics* **26**, 334.
- FLYNN G. J. and SUTTON S. R. (1992a) Element abundances in stratospheric cosmic dust: Indications for a new chemical type of chondritic material. *Lunar Planet. Sci.* **XXIII**, 373–374.
- FLYNN G. J. and SUTTON S. R. (1992b) Trace elements in chondritic stratospheric particles: Zinc depletion as a possible indicator of atmospheric entry heating. *Proc. 22nd Lunar Planet. Sci. Conf.*, 171–184.
- FLYNN G. J. and SUTTON S. R. (1992c) Trace elements in chondritic cosmic dust: Volatile correlation with Ca abundance. *Meteoritics* **27**, 220–221.
- FREDRIKSSON K. and KERRIDGE J. F. (1988) Carbonates and sulfates in CI chondrites: Formation by aqueous activity on the parent body. *Meteoritics* **23**, 35–44.
- FREDRIKSSON K., JAROSEWICH E., BEAUCHAMP R., and KERRIDGE J. (1980) Sulphate veins, carbonates, limonite and magnetite: Evidence on the late geochemistry of the C-1 regoliths. *Meteoritics* **15**, 291–292.
- GADSDEN M. (1968) Sodium in the upper atmosphere: Meteoric origin. *J. Atmos. Terrest. Phys.* **30**, 151–161.
- GOVINDARAJU K. (1984) 1984 compilation of working values and sample description for 170 international reference samples of mainly silicate rocks and minerals. *Geostand. Newsl.* **8** (Spec. Issue).
- GRADY M. M., GRAHAM A. L., BARBER D. J., AYLMEYER D., KURAT G., NTAFLLOS T., OTT U., and PALME H. (1986) Yamato 82042: An unusual carbonaceous chondrite with CM affinities. *Mem. NIPR Spec. Iss.* **46**, 162–178.
- GRAHAM A. L. and KURAT G. (1991) Phyllosilicates in the Yamato 82042 carbonaceous chondrite—primitive or not? *Lunar Planet. Sci.* **XXII**, 475.
- HOINKES G. and KURAT G. (1975) Preliminary report on the Bali carbonaceous chondrite. *Meteoritics* **19**, 416–417.
- HUGHES D. W. (1978) Meteors. In *Cosmic Dust* (ed. J. A. M. McDONNELL), pp. 123–185. Wiley.
- HUNTEN D. M., TURCO R. P., and TOON O. B. (1980) Smoke and dust particles of meteoric origin in the mesosphere and stratosphere. *J. Atmos. Sci.* **37**, 1342–1357.
- HUTCHISON R. (1987) Chromian-manganian augite in the interchondrule matrix of the Tieschitz (H3) chondritic meteorite. *Mineral. Mag.* **51**, 311–316.
- IVANOV A. V. (1989) The Kaidun meteorite: Composition and history. *Geochem. Intl.* **26**, 84–91.
- IWAHASHI J., YOSHIDA M., MIONO S., SANTOSH G., and SANTOSH M. (1991) Magnetic microspherules in Permian and Triassic bedded chert from Southwest Japan. *Proc. NIPR Symp. Antarct. Meteorites* **4**, 420–435.
- JAROSEWICH E., CLARKE R. S., JR., and BARROWS J. N. (1987) The Allende Meteorite Reference Sample. *Smithsonian Contrib. Earth Sci.* **27**, 49 p.
- JESSBERGER E. K., BOHSUNG J., CHAKAVEH S., and TRAXEL K. (1992) The volatile element enrichment of chondritic interplanetary dust particles. *Earth Planet. Sci. Lett.* **112**, 91–99.
- JOHNSON C. A. and PRINZ M. (1993) Carbonates in CM chondrites: Comparison with the CI group and implications for aqueous alteration. *Geochim. Cosmochim. Acta* **57**, 2843–2852.
- KALLEMEYN G. W., RUBIN A. E., and WASSON J. T. (1991) The compositional classification of chondrites: V. The Karoonda (CK) group of carbonaceous chondrites. *Geochim. Cosmochim. Acta* **55**, 881–892.
- KANE T. J. and GARDNER C. S. (1993) Lidar observations of the meteoric deposition of mesospheric metals. *Science* **259**, 1297–1300.
- KELLER L. P., THOMAS K. L., and MCKAY D. S. (1992) An interplanetary dust particle with links to CI chondrites. *Geochim. Cosmochim. Acta* **56**, 1409–1412.
- KLÖCK W., THOMAS K. L., MCKAY D. S., and PALME H. (1989) Unusual olivine and pyroxene composition in interplanetary dust and unequilibrated ordinary chondrites. *Nature* **339**, 126–128.
- KLÖCK W., BECKERLING W., SPETTEL B., FLYNN G., and SUTTON S. (1992a) Bulk composition and mineralogy of Antarctic micrometeorites. *Lunar Planet. Sci.* **XXIII**, 697–698.
- KLÖCK W., FLYNN G. J., SUTTON S. R., and NIER A. O. (1992b) Mineralogy of IDPs with known 4He and trace element contents. *Meteoritics* **27**, 243–244.
- KOEBERL C. (1993) Instrumental neutron activation analysis of geochemical and cosmochemical samples—a fast and reliable method

- for small sample analysis. *J. Radioanal. Nucl. Chem. Art.* **168**, 47–60.
- KOEBERL C., KURAT G., PRESPEER T., BRANDSTÄTTER F., and MAURETTE M. (1992) Bulk major and trace element analyses of unmelted micrometeorites from Cap Prudhomme, Antarctica. *Lunar Planet. Sci. XXIII*, 709–710.
- KORNBLUM J. J. (1969) Micrometeoroid interaction with the atmosphere. *J. Geophys. Res.* **74**, 1893–1907.
- KURAT G. (1975) Der kohlige Chondrit Lancé: Eine petrologische Analyse der komplexen Genese eines Chondriten. *Tschermaks Mineral. Petrol. Mitt.* **22**, 38–78.
- KURAT G. (1988) Primitive meteorites: An attempt towards unification. *Phil. Trans. R. Soc. London A* **325**, 459–482.
- KURAT G., MAYR M., NTAFLIS T., and GRAHAM A. L. (1989a) Isolated olivines in the Yamato 82042 CM2 chondrite: The tracing of major condensation events in the solar nebula. *Meteoritics* **24**, 35–42.
- KURAT G., PALME H., BRANDSTÄTTER F., and HUTH J. (1989b) Allende xenolith AF: undisturbed record of condensation and aggregation of matter in the solar nebula. *Z. Naturforsch.* **44a**, 988–1004.
- KURAT G., ZINNER E., and PALME H. (1989c) Primitive olivines with high trace element contents in Allende-AF aggregates. *Meteoritics* **24**, 290.
- KURAT G., BRANDSTÄTTER F., PALME H., SPETTEL B., and PRINZ M. (1991) Maralinga (CK4): Record of highly oxidizing nebular conditions. *Meteoritics* **26**, 360.
- KURAT G., KOEBERL C., PRESPEER T., BRANDSTÄTTER F., and MAURETTE M. (1992a) Bulk compositions of Antarctic micrometeorites: Nebular and terrestrial signatures. *Meteoritics* **27**, 246.
- KURAT G., PRESPEER T., BRANDSTÄTTER F., MAURETTE M., and KOEBERL C. (1992b) CI-like micrometeorites from Cap Prudhomme, Antarctica. *Lunar Planet. Sci. XXIII*, 747–748.
- KURAT G., BRANDSTÄTTER F., PRESPEER T., KOEBERL C., and MAURETTE M. (1993) Micrometeorites. *Geologia i Geofizika* **34**, 148–164 (in Russian).
- KYTE F. T. and SMIT J. (1985) Cretaceous-Tertiary spinels: High-temperature relicts from a major accretionary event. *Lunar Planet. Sci. XVI*, 473–474.
- KYTE F. T. and WASSON J. T. (1986) Accretion rate of extraterrestrial matter: iridium deposited 33 to 67 million years ago. *Science* **232**, 1225–1229.
- LEE M. S., RUBIN A. E., and WASSON J. T. (1992) Origin of metallic Fe-Ni in Renazzo and related chondrites. *Geochim. Cosmochim. Acta* **56**, 2521–2533.
- LOVE S. G. and BROWNLEE D. E. (1991) Heating and thermal transformation of micrometeoroids entering the Earth's atmosphere. *Icarus* **89**, 26–43.
- LOVE S. G. and BROWNLEE D. E. (1993) A direct measurement of the terrestrial mass accretion rate of cosmic dust. *Science* **262**, 550–553.
- MACKINNON I. D. R. and RIETMEIJER F. J. M. (1984) Bismuth in interplanetary dust. *Nature* **311**, 135–138.
- MAURETTE M., HAMMER C., BROWNLEE D. E., REEH N., and THOMSEN H. H. (1986) Placers of cosmic dust in the Blue Ice Lakes of Greenland. *Science* **233**, 869–872.
- MAURETTE M., JEIANNI C., ROBIN E., and HAMMER C. (1987) Characteristics and mass distribution of extraterrestrial dust from the Greenland ice cap. *Nature* **328**, 699–702.
- MAURETTE M., BROWNLEE D. E., and SCHRAMM L. S. (1989a) Giant micrometeorites from Antarctic blue ice. *Lunar Planet. Sci. XX*, 636–637.
- MAURETTE M., POURCHET M., BONNY P., DEANGELIS M., and SIRY P. (1989b) A new collection of micrometeorites, extracted from 100 tons of artificially melted blue ice, near Cap-Prudhomme in Antarctica. *Lunar Planet. Sci. XX*, 644–645.
- MAURETTE M., OLINGER C., WALKER R., and HOHENBERG C. (1989c) Noble gas measurements of extraterrestrial particles from polar sediments. *Lunar Planet. Sci. XX*, 640–641.
- MAURETTE M., OLINGER C., CHRISTOPHE MICHEL-LEVY M., KURAT G., POURCHET M., BRANDSTÄTTER F., and BOUROT-DENISE M. (1991) A collection of diverse micrometeorites recovered from 100 tonnes of Antarctic blue ice. *Nature* **351**, 44–47.
- MAURETTE M., IMMEL G., PERREAU M., POURCHET M., VINCENT C., and KURAT G. (1992a) The 1991 EUROMET collection of micrometeorites at Cap Prudhomme, Antarctica: discussion of possible collection biases. *Lunar Planet. Sci. XXIII*, 859–860.
- MAURETTE M., KURAT G., PRESPEER T., BRANDSTÄTTER F., and PERREAU M. (1992b) Possible causes of depletion and enrichment of minor elements in Antarctic micrometeorites. *Lunar Planet. Sci. XXIII*, 861–862.
- MAURETTE M., BROWNLEE D. E., JOSWIAK D. J., and SUTTON S. R. (1992c) Antarctic micrometeorites smaller than 50  $\mu\text{m}$ . *Lunar Planet. Sci. XXIII*, 857–858.
- MAURETTE M., KURAT G., PERREAU M., and ENGRAND C. (1993) Microanalyses of Cap-Prudhomme Antarctic micrometeorites. *Microbeam Anal.* **2**, 239–251.
- MÜLLER W. F., KURAT G., and KRACHER A. (1979) Chemical and crystallographic study of cronstedtite in the matrix of the Cochabamba (CM2) carbonaceous chondrite. *Tschermaks Mineral. Petrol. Mitt.* **26**, 293–304.
- MUTCH T. A. (1966) Abundances of magnetic spherules in Silurian and Permian salt samples. *Earth Planet. Sci. Lett.* **1**, 325–329.
- NAZAROV M. A., BRANDSTÄTTER F., and KURAT G. (1993) Carbonaceous xenoliths from the Erevan howardite. *Lunar Planet. Sci. XXIV*, 1053–1054.
- NELEN J., KURAT G., and FREDRIKSSON K. (1975) The Renazzo chondrite—a reevaluation. *Meteoritics* **10**, 464–465.
- PALME H., SUESS H. E., and ZEH H. D. (1981) Abundances of the elements in the solar system. In *Landoldt-Boernstein* (ed. K. SCHEIFERS and H. H. VOIGT), pp. 257–273. Springer-Verlag.
- PALME H., KURAT G., SPETTEL B., and BURGHELE A. (1989) Chemical composition of an unusual xenolith of the Allende meteorite. *Z. Naturforsch.* **44a**, 1005–1014.
- PERNICKA E., KURAT G., BRANDSTÄTTER F., and HERRWERTH I. (1985) Chainpur (LL-3): fractionated siderophile elements in chondrules, fragments, and chondrite matrix. *Meteoritics* **20**, 729–730.
- PERNICKA E., BAJT S., TRAXEL K., KURAT G., and BRANDSTÄTTER F. (1989) Composition and interrelationship of chondrules, lithic fragments and fine-grained matrix from Chainpur (LL-3). *Meteoritics* **24**, 316.
- PERREAU M., MAURETTE M., KURAT G., and ENGRAND C. (1992) Carbon-rich phases in Cap-Prudhomme micrometeorites. *Meteoritics* **26**, 274.
- PERREAU M., ENGRAND C., MAURETTE M., KURAT G., and PRESPEER T. (1993) C/O atomic ratios in micrometer-size crushed grains from Antarctic micrometeorites and two carbonaceous meteorites. *Lunar Planet. Sci. XXIV*, 1125–1126.
- PRESPEER T. (1993) Pauschalzusammensetzung und Mineralchemie von arktischen kosmischen Kügelchen und antarktischen Mikrometeoriten. Ph.D. thesis, Univ. Mainz.
- PRESPEER T., KURAT G., and MAURETTE M. (1992) Preliminary report on the composition of anhydrous primary phases in micrometeorites from Cap Prudhomme, Antarctica. *Meteoritics* **26**, 278.
- PRESPEER T., KURAT G., KOEBERL C., PALME H., and MAURETTE M. (1993) Elemental depletions in Antarctic micrometeorites and Arctic cosmic spherules: Comparison and relationships. *Lunar Planet. Sci. XXIV*, 1177–1178.
- RIETMEIJER F. J. M. (1985) Low-temperature aqueous and hydrothermal activity in a proto-planetary body: Goethite, opal-CT, gibbsite, and anatase in chondritic porous aggregate W7029\*A. *Lunar Planet. Sci. XVI*, 696–697.
- RIETMEIJER F. J. M. (1992) Interplanetary dust particle L2005T12 directly linked to type CM chondrite petrogenesis. *Lunar Planet. Sci. XXIII*, 1153–1154.
- RIETMEIJER F. J. M. and MACKINNON I. D. R. (1984a) Layered silicates in chondritic porous aggregate W7029\*A: a case of primary growth. *Lunar Planet. Sci. XV*, 687–688.
- RIETMEIJER F. J. M. and MACKINNON I. D. R. (1984b) Diagenesis in interplanetary dust: chondritic porous aggregate W7029\*A. *Meteoritics* **19**, 301.
- RIETMEIJER F. J. M. and MACKINNON I. D. R. (1985) Layer silicates in a chondritic porous interplanetary dust particle. *J. Geophys. Res.* **90**, D149–D155.
- RIETMEIJER F. J. M. and MACKINNON I. D. R. (1990) Titanium oxide Magneli phases in four chondritic porous interplanetary dust particles. *Proc. 20th Lunar Planet. Sci. Conf.* 323–333.

- RIETMEIJER F. J. M. and MCKAY D. S. (1986) Fine-grained silicates in chondritic interplanetary dust particles are evidence for annealing in the early solar system. *Lunar Planet. Sci. XVII*, 710–711.
- ROBIN E., CHRISTOPHE MICHEL-LEVY M., BOUROT-DENISE M., and JEHANNO C. (1990) Crystalline micrometeorites from Greenland blue lakes: Their chemical composition, mineralogy and possible origin. *Earth Planet. Sci. Lett.* **97**, 162–176.
- ROBIN E., BONTE P., FROGET L., JEHANNO C., and ROCCHIA R. (1992) Formation of spinels in cosmic objects during atmospheric entry: A clue to the Cretaceous-Tertiary boundary event. *Earth Planet. Sci. Lett.* **108**, 181–190.
- SCHRAMM L. S., BROWNLEE D. E., and WHEELLOCK M. M. (1989) Major element composition of stratospheric micrometeorites. *Meteoritics* **24**, 99–112.
- SCOTT E. R. D., BARBER D. J., ALEXANDER C. M., HUTCHISON R., and PECK J. A. (1988) Primitive material surviving in chondrites: Matrix. In *Meteorites and the Early Solar System* (ed. J. F. KERRIDGE and M. W. MATTHEWS), pp. 718–745. Univ. Arizona Press.
- STEELE I. M. (1986) Compositions and textures of relic forsterite in carbonaceous and unequilibrated ordinary chondrites. *Geochim. Cosmochim. Acta* **50**, 1379–1395.
- STEELE I. M. (1992) Olivine in Antarctic micrometeorites: Comparison with other extraterrestrial olivine. *Geochim. Cosmochim. Acta* **56**, 2923–2929.
- STEINWEG A., KRANKOWSKY D., LÄMMERZAHN P., and ANWEILER B. (1992) Metal ion layers in the auroral E-region measured by mass spectrometers. *J. Atmos. Terr. Phys.* **54**, 703–714.
- SUTTON S. R. and FLYNN G. J. (1988) Stratospheric particles: synchrotron X-ray fluorescence determination of trace element contents. *Proc. 18th Lunar Planet. Sci. Conf.*, 607–614.
- SUTTON S. R. and FLYNN G. J. (1989) Trace element compositions of interplanetary dust and terrestrial particles collected from the stratosphere. *Lunar Planet. Sci. XX*, 1091–1092.
- THOMAS K. L., ZOLENSKY M. E., KLÖCK W., and MCKAY D. S. (1990) Mineralogical descriptions of eight hydrated interplanetary dust particles and their relationship to chondrite matrix. *Lunar Planet. Sci. XXI*, 1250–1251.
- THOMAS K. L., KELLER L. P., FLYNN G. J., SUTTON S. R., TAKATORI K., and MCKAY D. S. (1992) Bulk compositions, mineralogy, and trace element abundances of six interplanetary dust particles. *Lunar Planet. Sci. XXIII*, 1427–1428.
- VAN DER STAP C. C. A. H., VIS R. D., and VERHEUL H. (1986) Interplanetary dust: arguments in favour of a late stage nebular origin of the chondritic aggregates. *Lunar Planet. Sci. XVII*, 1013–1014.
- WASSON J. T. (1985) *Meteorites. Their Record of Early Solar-System History*. W. H. Freeman.
- WASSON J. T. and KALLEMEYN G. W. (1988) Composition of chondrites. *Phil. Trans. R. Soc. London A* **325**, 535–544.
- WEISBERG M. K., PRINZ M., CLAYTON R. N., and MAYEDA T. K. (1993) The CR (Renazzo-type) carbonaceous chondrite group and its implications. *Geochim. Cosmochim. Acta* **57**, 1567–1586.
- ZOLENSKY M. E. (1987) Tochilinite in C2 carbonaceous chondrites: a review with suggestions. *Lunar Planet. Sci. XVIII*, 1132–1133.
- ZOLENSKY M. and LINDSTROM D. (1991) Mineralogy of 12 large “chondritic” interplanetary dust particles. *Lunar Planet. Sci. XXII*, 1557–1558.
- ZOLENSKY M. and MCSWEEN H. Y. (1988) Aqueous Alteration. In *Meteorites and the Early Solar System* (ed. J. F. KERRIDGE and M. S. MATTHEWS), pp. 114–143. Univ. Arizona Press.
- ZOLENSKY M., BARRETT R., and BROWNING L. (1993) Mineralogy and composition of matrix and chondrule rims in carbonaceous chondrites. *Geochim. Cosmochim. Acta* **57**, 3123–3148.

## APPENDIX: SUMMARY OF PETROLOGICAL OBSERVATIONS ON ANTARCTIC MICROMETEORITES

Object	M1	M2	M3	M4	M5	M6	M7	AM1
Group	Crystalline	Scoria	Phyllosilicate	Scoria	Scoria	Crystalline	Crystalline	Crystalline
Texture	Porphyritic	Aphan Highly vesicular Anhydrous relics	Aphan, vesicular Transitional to scoria Anhydrous relics	Quench Transitional to spherule Anhydrous relics	Vesicular, quench Transitional to spherule	Porphyritic Abundant vesicles	Poikilitic Abundant vesicles	Porphyritic Abundant vesicles
Phases	Opx (Fs20-23) Ol (Fa2-4,Al,Gr,Ca) Cpx Chr (67% Cr) Sulf (Fe) Metal (40% Ni) Glass	Ol (Fa23-31,very Al,Ti,Ca,Ni-rich) Glass Sulf Mt	Opx (Fs8,platy) Phyllo (metamorphic?) Mt Glass	Ol (Fa1,28) Glass Mt	Glass Mt Ol? Sulf	Ol(Fa19) Opx(Fs17) Chr FeNi-sulf Glass	Ol (Fa28-30,Ni) Plag Chr Mt (Ni)	Ol (Fa17) Aug Glass Chr
Surface	Irregular Granular Glass coat Rare thin mt coat	Irregular Rounded Vuggy Common mt coat	Irregular Glass coat Partial mt coat	Irregular Rounded Pitted Thick mt coat	Irregular Rounded Smooth Thick mt coat	Irregular Rounded Smooth Small pits Tiny COPS	Irregular Angular Thin mt coat in places	Irregular Smooth No mt coat
Comment	Fo welded to px- rich MM	Ol similar to All-AF	Condensation texture of opx? Fe metasomatism	Ol relics very different Matrix sample?	Na,K,Cl,Si-rich phases in vesicles	Voids weathering of sulf?	Voids weathering of sulf?	Voids weathering of sulf?



## APPENDIX: SUMMARY OF PETROLOGICAL OBSERVATIONS ON ANTARCTIC MICROMETEORITES (CONTINUED)

Object	3M5	3M6	3M7	3M8	4M1	4M5	4M12
Group	Scoria	Scoria	Phyllosilicate	Phyllosilicate	Phyllosilicate	Scoria	Scoria
Texture	Aphan Highly vesicular Anhydrous relics	Aphan Highly vesicular Anhydrous relics	Dense, scoriaceous rim Abundant voids Part of aggregate	Dense, fine-grained Voids + cracks	Dense, fine-grained Voids + cracks	Aphan Highly vesicular Anhydrous relics	Aphan Highly vesicular Phyllo relics?
Phases	Opx (Fs2) Glass Mt	Opx (Fs1) Glass Mt	Phyllo Glass Mt	Phyllo Laihumite (1.9% Ni) Glass Mt	Opx (Fs1.5) Phyllo Mt	Ol (Fa1.5) Opx (Fs2-3) Fe-oxide Glass Mt	Glass Mt Fe-oxide Phyllo (metamorphic)
Surface	Irregular Mt+COPS(?) coat	Irregular Mt coat	Highly irregular Part mt coat Part breccia	Angular Mt+Fe-oxide coat	Irregular Mt+Fe-oxide coat	Irregular Mt+COPS(?) coat C-rich contaminant	Irregular Mt coat
Comment	Fe metasomatism	Broken into two similar pieces	Aggregate of phyllo and terrestrial phases K-fsp, C, Fe-oxide	CM matrix sample?	Cracks emptied sulf veins?	Fe metasomatism Thick mt coat	Fe metasomatism

## APPENDIX: SUMMARY OF PETROLOGICAL OBSERVATIONS ON ANTARCTIC MICROMETEORITES (CONTINUED)

Object	AM2	AM4	AM5	AM8	AM9	AM10	3M2	3M4
Group	Scoria	Phyllosilicate	Phyllosilicate	Crystalline	Crystalline	Scoria	Scoria	Crystalline
Texture	Transition to crystalline; VL anhydrous relic Abundant open cracks	Dense, fine-grained; metamorphic Some voids, cracks	Dense, fine-grained Voids Scoriaceous cover	Porphyry Large voids	Porphyry Large vesicles	Aphan Highly vesicular	Porous aggregate Crystalline + Scoria Large voids	Poikilitic
Phases	Ol (Fa1, Cr, Mn) Mt (Ni)	Phyllo Dehydrated phyllo	Phyllo (Ni) Cronstedtite (Ni) Tochil Mt Glass	Ol Opx Mt Sp	Ol (Fa32-37, Ca, Cr) Aug Plag (An12) Glass Chr Sulf (11% Ni)	Glass Mt (primary and secondary)	Opx (Fs3-5) Aug Plag (An25) Fe-Oxide (1.4% Ni) Metal (5% Ni)	Opx (Fs3) Ol (Fa2, Cr) Tochil (3.5% Ni) Mt
Surface	Irregular Smooth Thick mt coat	Irregular Mt coat	Irregular Mt coat	Irregular Mt coat	Irregular Mt coat	Two irregular drops Mt coat	Irregular Mt + COPS coat Cl, pb-containing spots	Irregular/angular En poikilitic with round ol + Fe-ox
Comment	Soluble phase lost from cracks?	Al <sub>2</sub> O <sub>3</sub> at surface; Fe metasomatism; Phase lost?	GM mineralogy Phase lost?	Al <sub>2</sub> O <sub>3</sub> at surface	SiO <sub>2</sub> -rich glass at surface Phase lost?	Fe metasomatism	Fe metasomatism Aggregate matrix COPS?; Rare mt coat	Metasomatized aggregate from CM?
	aphan = aphanitic aug = augite chr = chromite cpx = clinopyroxene		en = enstatite metam = metamorphic mt = magnetite ol = olivine		opx = orthopyroxene ox = oxide plag = plagioclase porphyry = porphyritic			phyllo = phyllosilicate sp = spinel sulf = sulfide tochil = tochilinite