

Properties of Interplanetary Dust: Information from Collected Samples

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Abstract. The properties of hundreds of interplanetary particles have been determined by direct laboratory analysis of recovered samples. The particles that span the 1 μm to 1 mm size range have been collected from the stratosphere, from polar ice, and from deep sea sediments. Typically, these particles are black, somewhat porous and have chondritic elemental compositions. They are rather complex mineral assemblages in that they are mixtures of very large numbers of sub-micrometer-sized components. While the data are not totally representative of small interplanetary meteoroids at 1 AU they provide significant insight into the common physical properties of meteoroids. These properties can be used as guidelines for analysis of spacecraft and astronomical observations and for modeling solar system dust as well as some circumstellar dust in systems around other stars.

I. INTRODUCTION

Knowledge of the properties of interplanetary dust particles (IDPs) is important for modeling the origin, evolution and nature of the meteoroid complex. It also plays an important role in understanding a variety of solar system processes including the formation and survival of the IRAS infrared dust bands, interpretation of the zodiacal light, processes in comet tails, *space weathering effects* on asteroids and alteration of surfaces of airless bodies. In a broader perspective, the properties of solar system dust also relate in important ways to circumstellar dust around other stars. This aspect has gained increasing importance with the discovery that dust systems around young and evolved stars are common and that many have dust systems analogous to the Sun's zodiacal dust complex. This is particularly true for the Vega-like systems where dust may be generated by comets in a manner similar to the Kuiper belt of the solar system (Weissman 1984). Particles are generated by larger bodies, their orbits

decay by Poynting-Robertson drag, and they are also destroyed by hypervelocity collisions. These processes also dominate the evolution of interplanetary dust in the solar system and it is likely that dust properties are similar in the solar system and other circumstellar dust systems, particularly where dust may be generated by comets. At a minimum, the solar system dust properties are a reasonable analogue for the properties of dust in other circumstellar systems. In the solar system it is possible to collect dust samples and study them in detail with a wide variety of modern laboratory instrumentation. This direct information provides important insight into the limits of properties of particles in many different environments.

The idealized goal of this review would be to evaluate the various physical, chemical, isotopic, and mineralogical properties of interplanetary dust in the solar system. Due to a variety of complications, this lofty goal is impossible to accomplish with existing or even future samples. There may be several types and populations of particles that vary with size, location, orbital parameters and time. What can be done is to describe samples collected at Earth, in a limited size range, and over a limited time interval. Due to selective effects of Earth capture, of atmospheric survival, of collection, and even of personal preferences of the researcher, the collected samples are only approximately representative of particles currently colliding with the Earth.

This review contains information from two different types of collections of interplanetary particles. One is collected from the stratosphere and the other from Antarctic and Greenland ice. Although all of the particles are micrometeorites (unmelted particles) and interplanetary dust (IDP), it is common usage to refer to the stratospheric particles as IDPs and the polar ice particles as MMs (micrometeorites). Due to contamination and collection limitations, most of the particles collected in the atmosphere are limited to the 5 μm to 25 μm diameter range. Most of these small particles survive atmospheric entry without melting and they are only moderately altered by their capture from space and residence on Earth. They are probably fairly representative of the 15 μm diameter particle population at 1 AU, although gravitational focusing effects will enhance the proportion of low speed asteroid particles (Flynn 1994; Kortenkamp and Dermott 1998). Much larger extraterrestrial particles have been collected from deep sea sediments and from Greenland and Antarctic ice and they provide direct information on the meteoroids in the 20 μm to 1 mm size range (Kurat et al. 1994a; Maurette et al. 1994). The larger particles ($> 50 \mu\text{m}$) are too rare to be effectively collected in the stratosphere and they must be recovered from surface deposits where they are affected by a range of alteration processes that are not yet fully understood. These large particles are typically more strongly heated during atmospheric entry and these effects selectively alter or destroy denser, fragile, or higher velocity particles. On the ground they also require some selection processes to be distinguished from terrestrial materials. Whereas nearly all 15 μm particles survive atmospheric entry and can be collected in the stratosphere with probably only mild alteration (besides some heating) in the stratosphere (Brownlee 1981, 1985;

Warren and Zolensky 1994), the larger particles collected from the Earth's surface are a more biased sampling. Although biased, they do uniquely provide samples of larger dust particles of the 200 μm size range that actually dominate the mass of meteoroids in the interplanetary medium and the annual influx of extraterrestrial material to Earth (Kyte and Wasson 1986; Brownlee 1997).

II. ANTARCTIC AND GREENLAND MICROMETEORITES

Large unmelted interplanetary dust particles (*micrometeorites* MMs) are available from Greenland ice (Maurette et al. 1986) and from Antarctica (Maurette et al. 1991, 1994, 1996). In Greenland, micrometeorites can be collected from "cryoconite", a dark sediment in melt water lakes, which consists of dust and cocoons of blue algae and siderobacteria. Cryoconite contains about 10 g/kg fine-grained sand and dust mostly of terrestrial origin and about 800 cosmic spherules and 200 unmelted to partially melted micrometeorites. In Antarctica, micrometeorites can be collected by intentionally melting blue ice. One ton of Antarctic blue ice contains about 100 cosmic spherules with diameters $> 50 \mu\text{m}$ and about 500 unmelted to partially melted MMs 50–400 μm in diameter. Thus, large amounts of unmelted and almost unaltered samples of the interplanetary dust particles which contribute most to the recent accretion rate on Earth are available for study.

II.A. Mineralogy and Petrography of MMs

Many micrometeorites have experienced alteration by frictional heating in the atmosphere. They are partially or totally melted and consist of foamy glass with variable amounts of unmelted phases (scoriaceous MMs, cosmic spherules, respectively, Figs. 1a–2a). Others have been thermally metamorphosed but not melted, and surprisingly many retained their pristine mineralogy (unmelted MMs).

The pristine mineralogy of micrometeorites is remarkably simple (e.g., Maurette et al. 1991, 1993, 1994; Kurat et al. 1992, 1993, 1994a,b). Major minerals are olivine, low-Ca pyroxene, magnetite, and hydrous Mg-Fe silicates (phyllosilicates) like serpentine and saponite. The typical unmelted MMs are dense and low-porosity mixtures of anhydrous and hydrous phases in proportions ranging from all anhydrous (coarse-grained "crystalline" micrometeorites, Fig. 2b) to totally hydrous mineral assemblages (phyllosilicate micrometeorites, Fig. 3). Minor phases are Ca-rich pyroxenes, feldspars, Fe-Ni sulfides and metal, Mg-Fe hydroxides, MgAl and Fe-Cr spinels, perovskite, ilmenite, hibonite, and others (Kurat et al. 1994c; Beckerling and Bischoff 1995; Hoppe et al. 1995). A summary on the mineralogy, chemistry, and oxygen isotopic composition is given by Greshake et al. (1996). The major anhydrous silicates have variable

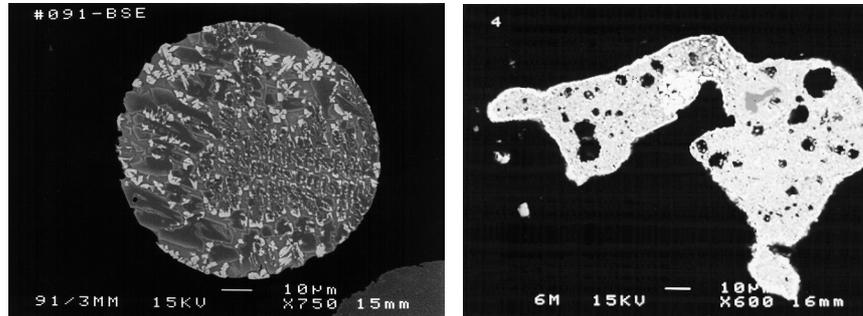


Figure 1a (left). Cosmic spherule from Antarctica, polished section, back scattered electron (BSE) scanning image, quench texture with olivine (dark gray to gray, compositionally zoned) and magnetite (white) in glassy matrix (gray). From mount 91/3 (particle 91).

Figure 1b (right). Scoriaceous micrometeorite 6M4 from Antarctica. BSE image of polished section shows magnetite-rich melt with vesicles (black) and an irregularly shaped relic olivine (gray). Note the magnetite cover.

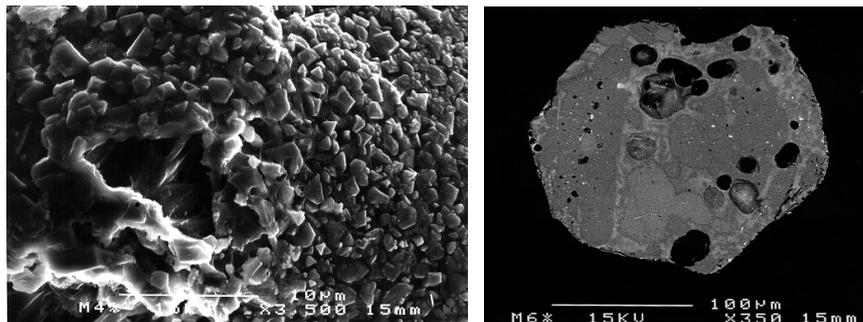


Figure 2a (left). Magnetite cover on micrometeorite M4 (Kurat et al. 1994) from Antarctica. Most thermally altered MMs are covered by magnetite.

Figure 2b (right). Crystalline micrometeorite M6 (Kurat et al. 1994a) from Antarctica. BSE image of polished section showing olivine (dark gray), pyroxene (gray) and dispersed sulfides and chromite (white). Such anhydrous mineral assemblages resemble olivine and olivine pyroxene aggregates in carbonaceous chondrites.

Fe/Mg ratios (unequilibrated mineral assemblage, Walter et al. 1994) and are usually very rich in minor elements as compared to their terrestrial counterparts and olivines from the most common meteorites, the ordinary chondrites. Ordinary chondrite matter is very rare among MMs and comprises less than 1 % by number (Walter et al. 1995). The phyllosilicates contain some elements in chondritic abundances (e.g., Ti, Al, Cr, Na, and K). Refractory minerals like Mg-Al spinel and perovskite are strongly enriched in refractory trace elements (e.g., rare earth elements, Sc, Zr, Hf, etc.; Kurat et al. 1994b, 1994d; Hoppe et al. 1995) as compared to chondritic rocks. Mineralogy, mineral chemistry, and the presence of refractory minerals in MMs are similar to those of CM-type



Figure 3. Phyllosilicate MM from Antarctica consisting of clay minerals (dark) and magnetite (framboidal and platy). This texture and mineral association are similar to those of CI and CM carbonaceous chondrites. BSE image. From mount 91/3 (particle 108).

(Mighei-type) and CR-type (Renazzo-type) carbonaceous chondrites. However, some differences between MMs and CM/CR chondrites are the abundance of Ca-poor pyroxene in MMs (most CM chondrites do not contain such pyroxenes) and the lack of very Fe-poor olivines with high Al and Ca contents in MMs (they are common in CM and CR chondrites).

II.B. Major, Minor and Trace Element Chemistry of MMs

Phyllosilicate-rich MMs are characterized by chondritic bulk major and minor element abundances, except for Ca, Na, Ni, and S that are depleted with respect to CI (and CM/CR) chondrites (Fig. 4a). The lithophile trace element abundances in phyllosilicate-rich MMs are similar to those in CM chondrites (that in turn are similar to those in CR chondrites), except for K which is overabundant. The abundances of siderophile elements in MMs deviate from those in CI and CM/CR chondrites. Only the refractory elements Os and Ir and the volatile Se have abundances similar to those in CI and CM/CR chondrites. The common siderophile elements Ni and Co are depleted with respect to chondritic abundances and are also fractionated from each other; the Ni/Co ratio is non-chondritic. Iron is somewhat enriched over chondritic abundances as are Au and As. The depletion of Ni, Co, and S is probably due to terrestrial leaching of Ni-bearing Mg-Fe sulfates from MMs (Presper et al. 1993). Micrometeorites do not contain sulfates, which, however, are abundant in CM and CI chondrites. Similarly, the Ca depletion of MMs compared to CM chondrites is probably due to leaching of carbonates, minerals that are common in CM chondrites but absent in MMs. The enrichments in MMs of Au, As, and K over chondritic abundances very likely are due to terrestrial contamination. These three elements are strongly enriched in the terrestrial crust as compared to chondrites.

Scoriaceous MMs have trace element abundances very similar to those of phyllosilicate MMs (Fig. 4b), except for large depletions and re-enrichments

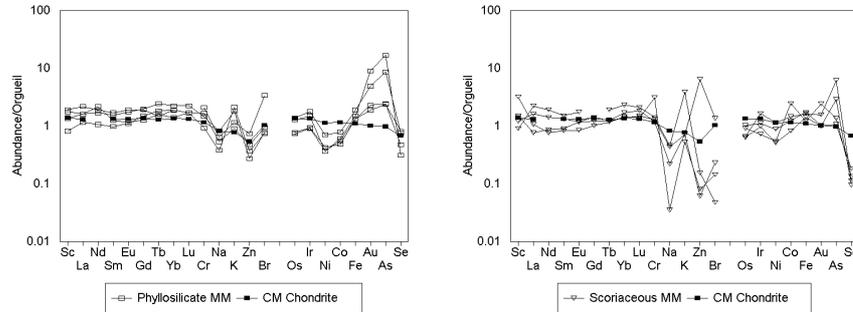


Figure 4a (left). CI-normalized selected element abundances in phyllosilicate MMs and CM chondrites (data from Kurat et al. 1994b; Anders and Grevesse 1989; Palme, pers. comm.). There is a close match between the compositions of MMs and CM chondrites for the refractory elements (Sc–Cr and Os, Ir). The siderophile element abundances are disturbed by loss of soluble minerals (sulfates containing Ni and Co) and by terrestrial contamination (Au and As).

Figure 4b (right). CI-normalized selected element abundances in scoriaceous MMs and CM chondrites (data sources as for Fig. 4a). The volatile elements (Na, K, Zn, Br, As and Se) are either depleted or enriched with respect to CI chondrites, a consequence of the heating event during atmospheric entry and subsequent contamination in the terrestrial environment.

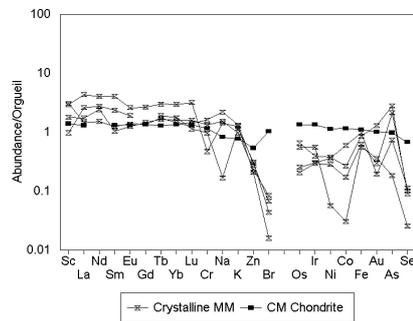


Figure 5. CI-normalized selected element abundances in coarse-grained crystalline MMs and CM chondrites (data sources as for Fig. 4a). Most fractionations with respect to CI chondrites are indigenous, except for the enrichment in As which is of terrestrial origin.

in volatile elements (Na, K, Zn, Br, As, Se). Clearly, scoriaceous MMs were formed from phyllosilicate meteoroids by heating during atmospheric entry.

Anhydrous crystalline MMs deviate in composition from chondrites, a feature typical also for anhydrous aggregates and chondrules in carbonaceous chondrites (Fig. 5). Elemental fractionations are usually not very severe for the refractory lithophile elements but are commonly strong for volatile and siderophile elements, a consequence of the conditions prevailing during aggregation.

Micrometeorites (and also IDPs) are surprisingly rich in carbon. Perreau et al. 1993 and Engrand et al. 1994 found C/O ratios that on average were higher than those in CI chondrites, the most C-rich chondrites. Micrometeorites are up to $5\times$ richer in C than CM/CR chondrites.

II.C. Isotope Abundances in MMs

Several elements have isotopic compositions that are non-terrestrial and, in some cases, also non-solar, in both, MMs and IDPs (e.g., McKeegan 1987a, b; Stadermann 1990; Kurat et al. 1994d; Hoppe et al. 1995; Engrand et al. 1996; see also below). Specifically, anomalies in the isotopic abundances of H, C, N, and O are common, similar to carbonaceous chondrites.

II.D. Rare Gas Abundances in MMs

Interplanetary dust has been exposed to the solar wind and cosmic rays for a sufficiently long time to accumulate large amounts of solar noble gases and spallogenic isotopes. Micrometeorites have, for example, very high He (up to $10^{-1} \text{ cm}^3 \text{ g}^{-1}$ STP) and Ne contents – in excess of $10^{-5} \text{ cm}^3 \text{ g}^{-1}$ STP – comparable only to a few very gas-rich chondrites and to lunar soil (Olinger et al. 1990; Maurette et al. 1991; Nier 1994). He and Ne isotope abundances are similar to those of solar energetic particles (SEP) and thus confirm the extraterrestrial origin of IDPs and MMs (and of some cosmic spherules). A minor contribution of cosmic ray spallation Ne was also identified. Interplanetary dust (IDPs and MMs) clearly was exposed to cosmic rays and to the solar wind. The particles must have been of the size as recovered while they were exposed to the solar wind. Therefore, MMs and IDPs (Nier and Schlutter 1990) are true interplanetary dust meteoroids and cannot be atmospheric break-up products of a larger-sized meteoroid.

II.E. Conclusions from MM Studies

Micrometeorites represent the main mass of extraterrestrial matter accreting onto the Earth today. Whereas meteorites, that deliver only a few percent of the total incoming mass on Earth, are dominated by ordinary chondrites, micrometeorites bear some similarities to the rare CI/CM/CR carbonaceous chondrites. However, they also differ from them in so many aspects that they have to be considered as a solar system matter of its own. The features of MMs that are different from those of chondrites are likely to be of primordial origin. These include the mineral abundances, mineral chemistry and the bulk C content.

There is considerable overlap in mineralogical and chemical composition between MMs and IDPs (e.g., phyllosilicate dominated particles, olivine and pyroxene abundances and chemical compositions) but the fluffy, fine-grained olivine aggregates and GEMS that are abundant in IDPs are not found so far in MMs. However, the particles of sizes $< 50 \mu\text{m}$ have not been investigated properly yet and that is where the crossover in mineralogical and petrographical features of MMs and IDPs can be expected. Deviations of the dust composition

from that of chondrites are very probably due to extraction of water-soluble sulfates and carbonates and to contamination in the terrestrial environment.

III. STRATOSPHERIC INTERPLANETARY DUST

III.A. Shape and External Morphology

The vast majority of the typically 15 μm sized IDPs have shapes that can be modeled by rather equidimensional forms. The exteriors of most particles can be roughly approximated by ellipsoids with ratios of maximum length to minimum width that are less than two. Typical particles are aggregates of smaller components and a good model of the overall shape is a lumpy ellipsoid with second order surface modulations in the size range from 0.1 μm to a few microns. The irregularities consist of both cavities and protrusions. Highly non-regular shapes such as plates and rods are essentially nonexistent for whole particles although some subcomponents do have these shapes. Only fairly uncommon particles have wedge-shaped angular structures generally seen as fragments produced by comminution of solid rocks. Some of the collected 15 μm particles are spherical but in most cases this is thought to be due to atmospheric melting, a process experienced by many particles larger than 100 μm . Typically, the surface texture is related to the interior structure and the porosity. The most common IDPs have roughly chondritic elemental compositions and, as shown in Fig. 6, they have surface properties ranging from smooth (CS or chondritic smooth) to those that are highly rough and porous (CP or chondritic porous).

The CS particles are smooth at the micrometer scale and usually have solid non-porous interiors. Many of the CS particles are dominated by hydrated silicates. Their smooth surfaces and the presence of hydrated silicates are related, in part, to aqueous alteration and compaction processes that modified the precursor materials. Many of the CS particles are mineralogically similar to the matrix of CM and CI chondrites and phyllosilicate micrometeorites. Not all particles, however, that contain hydrated minerals have smooth exteriors. Some are porous with rough exteriors and some are rough due to the matrix construction of masses of layer silicates. Most particles with rough surfaces at the micrometer scale are of the CP type and they are usually aggregates of large numbers of sub-micrometer grains and components. Most of these are dominated by anhydrous minerals but some are composed of hydrated phases. The fundamental building blocks of the porous particles are usually grains of 0.3 μm size. They form aggregate structures that range in porosity and degree of filling materials that assist in bonding them together. The surfaces of most porous particles are highly irregular and often contain deep cavities that in some cases provide open line-of-sight pathways that pass entirely through the particle. In general, the porous particles with rough surfaces are simply aggregates of small particles clustered in random open fashion with no apparent order. There is no evidence that sub-grains were ever linked end to end forming dendritic networks as is seen in smoke particles formed by condensation.

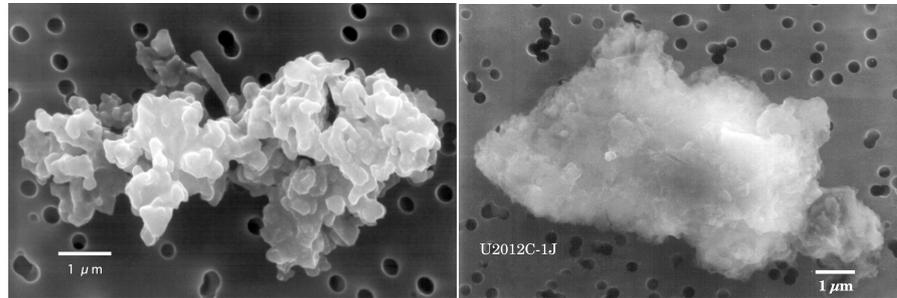


Figure 6. SEM images of (left) a porous chondritic IDP and (right) a smooth chondritic IDP.

Roughly 15 % of the IDPs are essentially single mineral grains or are simple assemblies of only a few mineral grains. Most commonly these are olivine, enstatite or pyrrhotite grains. These particles differ from the particles with chondritic composition in that they are usually more angular with broad planar surfaces. Most are anhedral although a few do show crystal faces.

III.B. Density

Particle density and area-to-mass ratio are important parameters for many processes involving interplanetary dust and it is common for calculations and models to assume a single nominal particle density. For real particles, however, there is a range of densities and even the meaning of their density is somewhat complex due to their irregular shapes. For non-spherical solid particles, the effective density will in most cases be less than the bulk density of the material that the particle is made of. For example, with higher average cross-section area, the Poynting Robertson drag force of irregular particles is higher than for spheres of the same mass and composition. The shape effects are particularly important for small particles because cavities and protrusions can make up an appreciable fraction of the particle diameter. When trying to determine the density there is often some ambiguity in what to use for its volume. In many cases, the volume can be well approximated by an imaginary rubber sheet that would surround the particle and contact its highest points.

The densities of particles can be measured in a variety of ways (Strait et al. 1995; Flynn and Sutton 1988; Fraundorf et al. 1982; Maetz et al. 1994). The most extensive study was the measurement of 150 chondritic 10 μm IDPs by Love et al. (1994). They estimated particle volumes by measuring the cross section of particles from SEM images and by measuring the heights by differential focus methods. Masses were measured by quantitative X-ray techniques that determined the total mass of elements heavier than oxygen. A weakness of the study was that C could not be measured and O had to be calculated by stoichiometry. This study found a range of 0.3 to 6.2 g cm^{-3} for chondritic composition particles with an average at 2.0 g cm^{-3} (Fig. 7). About 75 % of the measured densities were between 1 and 3 g cm^{-3} . Because the samples were collected in the stratosphere, the relative proportions of dense

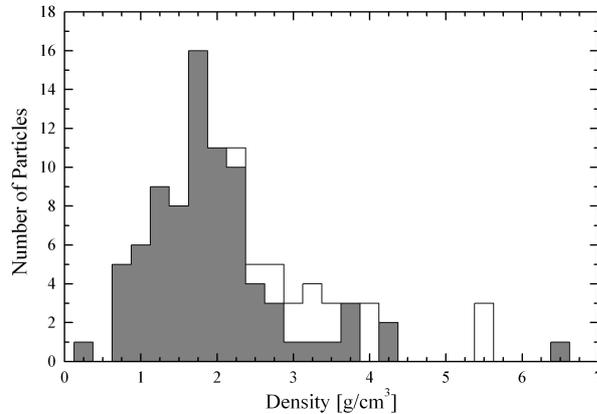


Figure 7. The density distribution of stratospheric IDPs measured by Love et al. (1994). The open bars refer to particles that are S-rich and contain abnormally large FeS grains.

and low-density particles are affected by the atmospheric fall speed. Dense particles fall at higher speeds and are relatively underrepresented. Attempting to correct for this effect gives a rough estimate of the mean density of particles impacting the top of the atmosphere of 2.8 g cm^{-3} , but the peak in the density distribution remains at 2.0 g cm^{-3} for $15 \mu\text{m}$ particles.

The Love et al. (1994) work covered only chondritic composition particles. It did not include non-chondritic particles such as those composed of single large mineral grains and it also did not include particles that fragmented during collection. The particles composed of single mineral grains are rare but they are usually non-porous and inclusion of these would shift the mean particle density to a slightly higher value. The most common dense particles are those dominated by FeS, which has a bulk density of 5 g cm^{-3} . Fragile particles that broke up during collection are perhaps the most porous IDPs and if their existence could be accurately accounted for, they would partly compensate for the also unmeasured solid mineral grain particles. It is significant that there are at least a few very highly porous IDPs in the stratospheric collections: it proves that they indeed do exist. They may be more abundant in space although it seems unlikely that they comprise a major fraction of $15 \mu\text{m}$ meteoroids.

Another method to study masses and densities of IDPs involves less assumptions, but is also more complicated, and consequently only fewer IDPs have been analyzed so far (Maetz et al. 1994; Arndt et al. 1996a): *Scanning Transmission Ion Microscopy* (STIM) measures the energy loss of few-MeV protons when they pass through matter. From the energy loss the mass to area ratio (area density) can directly be evaluated with only little dependence on the major element composition (Lefevre et al. 1987). By scanning a proton beam over the IDP one obtains the area density profile and thus also information on the porosity and on portions of the sample with varying density (Fig. 8a). With the very small proton beam sizes ($< 250 \text{ nm}$) available at the

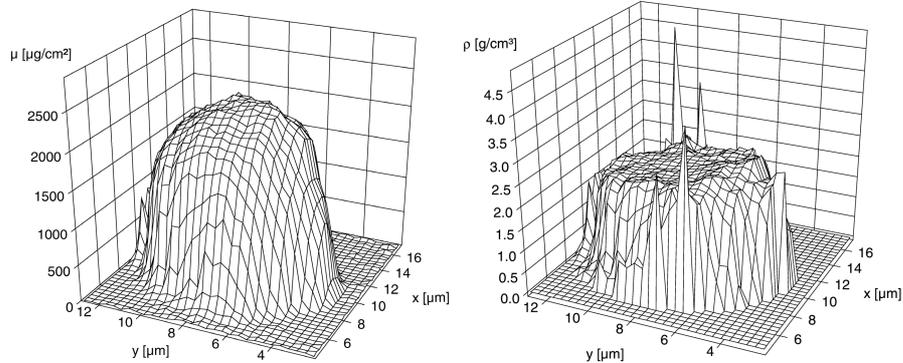


Figure 8a (left). Area density profile of IDP L2005B11 as obtained by scanning transmission ion spectroscopy.

Figure 8b (right). Density profile of IDP L2005B11 that results from combining the area density profile (Fig. 8a) and the figure of the IDP obtained from white light interferometry (resolution: $0.05 \mu\text{m}$). The mass of the particle is 1.2 ng , its density 2.3 g cm^{-3} .

new Heidelberg Proton Microprobe a lateral resolution of $1 \mu\text{m}$ for typical particle diameter of $10\text{--}15 \mu\text{m}$ can be achieved (Arndt et al. 1997). By integration over the particle area the total IDP mass can be calculated with a detection limit of about $1\text{--}2 \cdot 10^{-12} \text{ g}$ with 2 MeV protons (Maetz 1994). If the geometric surface profile and thus the volume of the IDP is known from, e.g., white light interference microscopy, both the mean density and the density profile can be calculated (Fig. 8b) (Maetz et al. 1994). Up to now this procedure resulted in the determination of four IDP masses ranging from 7.2 to 1191 pg and the densities of two IDPs, 1.0 and 2.3 g cm^{-3} (Maetz et al. 1994; Arndt et al. 1996a).

A remarkable result from particle density measurements is that common IDPs are much more porous than meteorites but they are not as highly porous as is sometimes assumed in modeling (Greenberg and Hage 1990; Greenberg and Gustafson 1981). The most common measured density is 2 g cm^{-3} consistent with moderate porosity near 40% . Particles with a density below 1 g cm^{-3} do exist but they are rare. The scarcity of very low density particles is consistent with the basic construction of common IDPs. Most of the porous particles are simply loose aggregates of rather equidimensional sub-micrometer components. In totally random aggregates of similar sized components it is difficult to have a porosity significantly higher than 50% because of the scarcity of contacts between grains to hold the particles together. As the density decreases the grain contacts along a transecting plane will reach zero and the particle becomes unbound. Microtome sections of very low density particles show very few grain-grain contacts in the section-plane and most of the grain-grain contacts that actually bind the particle together occur above and below the plane (Fig. 9).

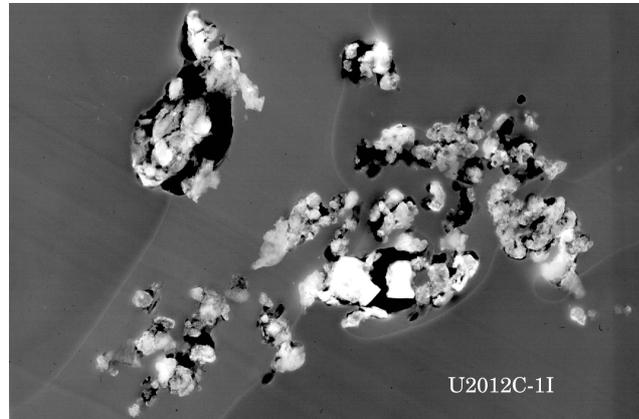


Figure 9. A TEM view of a $0.1 \mu\text{m}$ thin microtome section of a single porous IDP. The medium gray and black areas are epoxy and pore spaces, respectively, where material was torn out during cutting. The epoxy areas denote open space and show how unconnected the particle is on the section plane. Width of field is $10 \mu\text{m}$.

Densities much lower than 1 g cm^{-3} have been suggested for meteoroids (Ceplecha 1977), but this does not seem possible for $15 \mu\text{m}$ particles based on the structures observed in collected particles. To obtain such low densities would require a level of organization not seen in the samples. For example, extreme densities would require void spaces considerably larger than the $0.3 \mu\text{m}$ size of the building blocks of the aggregates. This in turn would require either thin sheets of a binding material such as carbonaceous organic or an organizational framework where the sub-grains would be organized into non-random chains to form beads-on-a-chain structures. It is also noteworthy that the mean densities for IDPs obtained in the laboratory roughly coincides with the mean densities estimated from in-situ measurements in comet Halley's coma (Maas et al. 1989)

Although densities of particles that are much larger or smaller than $15 \mu\text{m}$ have not been widely measured, it is possible to make predictions on these based on the assumption that they are composed of the loose aggregate structures of $15 \mu\text{m}$ IDPs. The aggregate structure of IDPs suggests that as particles approach sizes smaller than a micron they will primarily be solid objects and have densities $> 2 \text{ g cm}^{-3}$. For particles larger than $15 \mu\text{m}$ it is possible that increasingly larger void spaces will lead to increasing lower densities. If $100 \mu\text{m}$ particles are made by loose clustering of $15 \mu\text{m}$ particles, the larger particles will have lower densities. While it seems unlikely that an appreciable fraction of $15 \mu\text{m}$ IDPs can have densities $< 1.5 \text{ g cm}^{-3}$, it is possible that larger particles could have densities $< 1 \text{ g cm}^{-3}$. It is also possible that particles $> 100 \mu\text{m}$ are different from smaller particles either because they are derived from different sources or have undergone a different set of selection processes in the interplanetary medium that allow them to survive and reach 1 AU.

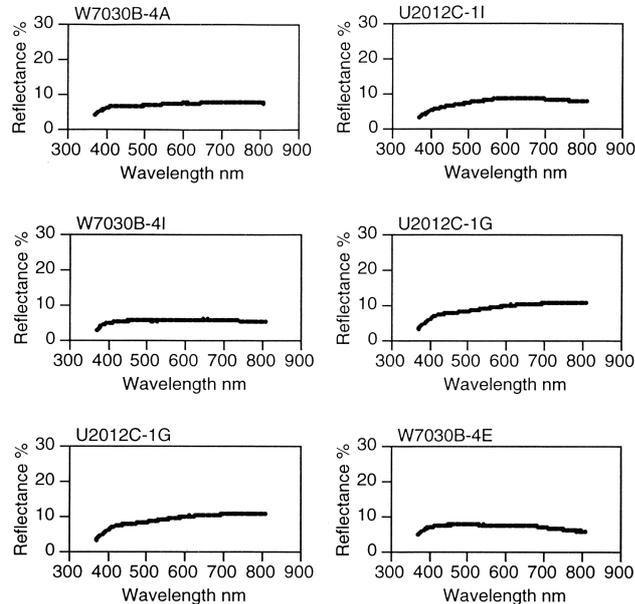


Figure 10. Reflectance spectra for individual $15 \mu\text{m}$ IDPs (Bradley et al. 1996).

At the same time, it is likewise possible that the much-larger-than-IDP particles represent the common meteorite constituents like chondrules and inclusions that are rarely found in IDPs.

III.C. Optical and Infrared Properties

In the optical microscope, most IDPs are opaque, fine-grained dark objects. Under close inspection, they are aggregate structures of mostly black irregular materials with occasional clear (silicate) grains and reflective (sulfide) grains ranging up to microns in size. Spectral reflectance measurements of individual stratospheric IDPs show that typical $15 \mu\text{m}$ IDPs have visible albedos in the range 5–15 % (Bradley et al. 1992; Bradley et al. 1996). Most have rather featureless reflectance curves except for downturns below 450 nm (Fig. 10). Most of the measured spectral reflectance curves resemble C type asteroids although some have increasing reflectance in the red, analogous to the P and D asteroid classes. The low albedo of typical IDPs is presumably due to their porous structure and the presence of strongly absorbing material. The absorbing materials include carbon, small sulfides and GEMS (sub-micrometer components composed of glass with embedded small metal and sulfide grains) (Bradley 1995). It is possible that the abundant nm-sized metal grains in GEMS are the major source of absorption in many of the IDPs. If this is the case, then the source of absorption in these particles would be very different from that in carbonaceous chondrites that are not reduced and do not contain abundant nanophase metal grains.

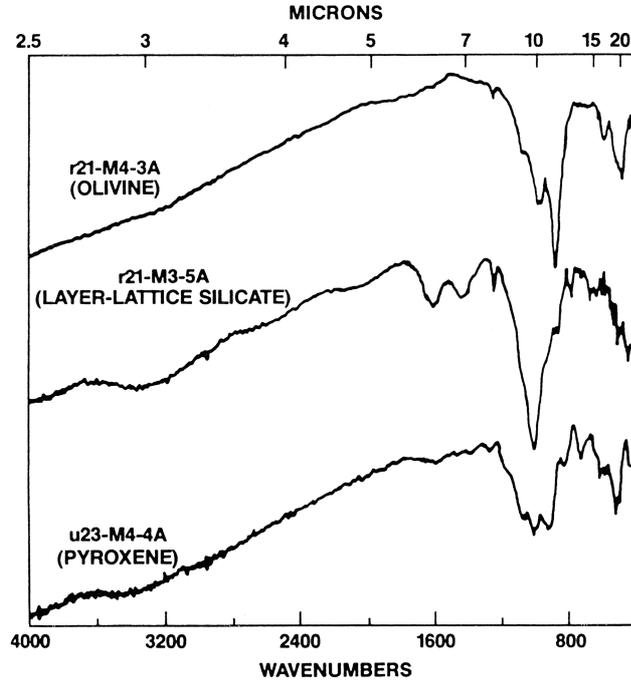


Figure 11. Infrared signatures of the three main infrared IDP groups showing differences in the 10 μm silicate feature (Sandford and Walker 1985).

The infrared properties of individual IDPs have been measured and they provide the basis of an IDP classification scheme (Sandford and Walker 1985). The shape of the 10 μm silicate absorption feature in the IR spectrum of particles varies in width, position, and fine structure (Fig. 11). These variations led to the establishment of the *olivine*, *pyroxene*, and *hydrated* subtypes. These classifications imply that the silicate feature of a particular IDP most closely matches the laboratory spectra of these minerals. Particles in any of the three classes can contain other minerals and although the IR classification is only a semi-quantitative assessment of a particles' bulk mineralogy, it is a very valuable classification scheme. The fine structure in the silicate structure also provides a useful means for comparing IDPs studied in the laboratory with dust in comets (cf. Sekanina et al., this volume) and dust around other stars. Fine structure producing a bump at 11.2 μm has been seen in several comets (Hanner et al. 1994) and has also been seen in dust around Beta Pictoris and related objects (Butner et al. 1994). This feature is attributed to the presence of olivine, although the best spectral match with IDPs is with the *pyroxene* class that is actually composed of glass with only minor amounts of crystalline silicate (Bradley et al. 1992). The *olivine* IDP class actually produces too much of a pronounced 11.2 μm fine structure relative to the cometary and circumstellar dust spectra. The similarity of the 11.2 μm fine structure in some IDPs, in comets, and Beta Pictoris and its apparent absence in normal

interstellar “astronomical silicate” provides important clues to processing of grains in nebular or circumstellar environments.

Recent infrared observation reveal the dominance of Fe-poor, Mg-rich sub-micron crystalline and glassy (or amorphous) olivine and pyroxene grains as the solid dust component of comet Hale-Bopp (Hanner et al. 1998). Such a paragenesis has previously been encountered only in anhydrous chondritic aggregate IDPs strengthening the link between this type of IPDs and comets (Hanner 1999) that is further supported by prevalence of Fe-poor, Mg-rich silicates in comet Halley’s dust (Jessberger 1999).

III.D. Classification and Mineralogy

Typical IDPs are fine-grained mixtures of thousands to millions of mineral grains and amorphous components. Like the most primitive meteorites, most of the IDPs are unequilibrated in the sense that a given phase in a particle can have wide ranging elemental compositions. The particles have not generally undergone metamorphic processing similar to those that equilibrated most ordinary chondrites. The particles are small, very complex and are difficult to fully characterize in a simple way. In some cases the groups may have a genetic meaning, in others, particles from different groups could be derived from common parents. Unfortunately, due to their small size, it is often difficult to classify them into such genetically meaningful groups. This can be attempted as was done with chondrites but it would be difficult using only 15 μm samples. The particles do, however, clearly fall into major categories and several investigators have suggested classification schemes (Brownlee et al. 1977; Sandford and Walker 1985; Bradley et al. 1988; Mackinnon and Rietmeijer 1987; Rietmeijer 1994). The following list is not very sophisticated but it does accommodate most unmelted particles. There are strongly heated to melted particles that are not included here because their original properties were significantly modified during atmospheric entry.

The great majority of the particles fall into the chondritic class and have approximately chondritic elemental composition as their identifying characteristic. Normally this requires that the abundant elements Mg, Al, Si, S, Ca, Cr, Mn, Fe, and Ni that can readily be detected by EDX are present within a factor of a few in CI chondritic abundance. It is often feared that this composition is used as a filter in identifying extraterrestrial particles and that thereby other particle classes are overlooked. This surely does happen, but it can only be a very minor effect. On good stratospheric collections, when the atmosphere is not contaminated by volcanic ash particles, the chondritic particles often comprise over 75 % of the total particles that are not Al-rich space debris or clear-cut contaminants from the aircraft. Most of the non-chondritic composition particles that are identified as extraterrestrial usually have some chondritic composition material attached to them. In most cases, they appear to be mineral grains or coarse grained assemblages of materials that were previously imbedded in fine-grained material similar to the chondritic groups. As such, they should not be thought of as necessarily genetically important

Table 1

Interplanetary Dust Types.

<i>Chondritic</i>	<i>Coarse Grained</i>
anhydrous (pyroxene, olivine)	sulfide
hydrated (layer silicate)	olivine
	pyroxene
	metal
	CAI
	carbonate
	phosphate

sub-classes. They are, however, legitimate interplanetary particles and their strength, density, and optical properties can differ significantly from normal chondritic particles.

Table 1 lists the coarse grained (non-chondritic) particle types in descending order of occurrence. The most common coarse grained particles are sulfides. These include particles that are single pyrrhotite crystals, polycrystalline mixtures of large sulfide grains (Zolensky and Thomas 1995), and large sulfide grains with various amounts of fine-grained chondritic material. Less common are olivine and pyroxene dominated particles (Zolensky and Barrett 1994) and the least common are particles dominated by phases that are abundant in (meteoritic) Ca-Al-rich inclusions (CAIs), large FeNi metal grains, carbonates, and phosphates.

The chondritic particle group is split into the hydrous and anhydrous classes. There is no perfect definition of the difference between the two groups. The hydrous class must have hydrated silicates but they also contain various amounts of anhydrous silicates. Even this definition is sometimes difficult to use with certain particles. These particles are typically identified by lattice fringe imaging of hydrous phases in the TEM. In some particles the “hydrous silicate” is very poorly ordered and difficult to image. This can be an original property but it can also result from thermal decomposition of layer silicates during atmospheric entry. The anhydrous particles are composed of anhydrous phases. Ideally, to be an official anhydrous IDP, it should have no hydrated phases but thermally decomposed or small amounts of hydrated phases are difficult to detect. There are also cases of what appear to be normal anhydrous IDPs that contain a 1 μm chunk of hydrated silicate (0.1 % of the IDP mass). It could be a tiny breccia fragment included into the IDP parent body, but at this level of “contamination,” it also could be a tiny piece of a hydrated IDP that was picked up from the collection substrate or during handling.

Three distinct mineralogical classes of chondritic IDPs are recognized. They are referred to as the *pyroxene*, *olivine*, and *layer silicate* classes. The dis-

tinct classes were first identified using infrared (IR) spectroscopy (Sandford and Walker 1985), and later confirmed using automated X-ray point-count analyses (Bradley 1988; Germani et al. 1990). The *pyroxene* and *olivine* classes are dominated by the anhydrous silicates pyroxene, olivine, (and glass), while the *layer silicate* class contain hydrated (layer lattice) silicates. Most chondritic IDPs belong to one of the three categories, although several exceptions have been reported. Examples include IDPs with approximately equal proportions of olivine and pyroxene, and others with both anhydrous and hydrated silicates.

The *pyroxene* IDPs are the most inherently interesting and intensively studied because they are highly porous, fluffy objects unlike any other known class of meteoritic materials. Most are complex admixtures of 0.1–5 μm diameter single mineral grains, glass, carbonaceous material, GEMS, and other fine-grained matrix material. The average grain size within the matrices is ≈ 100 nm or less. Enstatite is the most abundant of the single mineral grains and it occurs as euhedral crystals, platelets, and sometimes as whiskers (rods and ribbons). Some of the platelets, rods, and ribbons contain crystallographic defects suggestive of nebular gas-to-solid condensation (Bradley et al. 1983). Other enstatites (and forsterites) contain elevated Mn and Cr abundances consistent with gas-to-solid condensation (Klöck et al. 1989). Fe-rich sulfides are also important single mineral constituents as well as minor amounts of forsteritic olivine. Less abundant single-mineral species include alumino-silicate glass (Bradley 1994), metal (FeNi), and FeNi carbides (Christoffersen and Buseck 1983; Bradley et al. 1984).

The carbonaceous material in *pyroxene* IDPs is predominantly disordered (as opposed to graphitic). It is found throughout the IDPs both as discrete clumps and as a matrix with embedded mineral inclusions. Carbonaceous material enriched in ^{15}N has been observed in one IDP (Messenger et al. 1996), and large D/H anomalies are believed to be associated with the carbonaceous material (Messenger and Walker 1997). PAHs have been observed in several IDPs. It is likely that *pyroxene* IDPs contain other indigenous organics but, since the particles are pulse heated during atmospheric entry (but cf. Bonny et al. 1988), collected in silicone oil, and cleaned with solvents, the issue of indigenous organics is complicated.

In the *olivine* class, the IDPs typically are coarse-grained but some of them also contain carbonaceous material, glass, and fine-grained matrix components (Bradley et al. 1989; Christoffersen and Buseck 1986). In other words, there appears to be some mineralogical overlap between the *olivine* and *pyroxene* classes of IDPs. However, a conspicuously large number of *olivine* IDPs exhibit mineralogical evidence of strong heating (e.g., Fe-sulfides with magnetite rims) and they do not contain solar flare tracks. Thus, it is likely that the *olivine* class contains members that were severely heated during atmospheric entry. Laboratory experiments have shown that *layer silicate* IDPs transform into olivine-rich IDPs with heating (Greshake et al. 1998).

The relationship between the three classes of IDPs is unclear. Since solar flare tracks have been found in IDPs from each class, the mineralogical differ-

ences are believed to be indigenous rather than the effects of strong heating during atmospheric entry. However, tracks are commonly found in *pyroxene* and *layer lattice* silicate IDPs but they are conspicuously absent in most *olivine* class IDPs. Therefore, the *olivine* class may include a significant number of IDPs that were strongly heated (above 650°C) during atmospheric entry.

It is likely that at least some of the mineralogical differences between the three mineralogical classes reflect different types of parent bodies. For example, the *layer-lattice silicate* IDPs are probably from hydrous parent bodies where there was significant aqueous alteration. The mineralogy of several *layer silicate* IDPs is similar to the mineralogy of the CI and CM chondrites whose asteroidal origins are believed. Whether these differences are reflective of distinct classes of parent bodies is unknown. High-speed IDPs thought to be of cometary origin typically belong to the *pyroxene* and *olivine* classes, whereas most low-speed IDPs belong to the *layer-lattice silicate* class.

III.E. Elemental Composition

Given the large scale compositional difference in the solar system as a whole that reflects itself in density differences from the inner to the outer planets, given the compositional heterogeneity within the planets as best known from the Earth, and finally, given the complex and manifold diversity among primitive meteorites and their constituents, the elemental composition is one essential parameter to delineate the provenance and history of solar system material, including IDPs. Because of the possible sources of IDPs – they include asteroids also from the outer belt, or comets – the understanding of which probably is crucial for the understanding of the history of the solar system as a whole, quite some experimental efforts were made and highly sophisticated techniques were developed to obtain the elemental compositions of IDPs. This is difficult because of the small size of typically 15 μm and because each IDP has to be treated as an individual entity, like a whole meteorite.

For many purposes, it is sufficient to use a model composition of interplanetary dust that fairly closely matches chondritic elemental composition. A good approximation of typical IDP compositions is nominal chondritic composition for major element ratios with a match to CI chondrites (Anders and Grevesse 1989) within a factor of two. There is particle-to-particle variation and there are notable systematic variations from chondritic composition (e.g., Ca, C, and Zn as well as other volatile minor elements) but still a fairly monotonous chondritic composition is an excellent approximation for typical cases (Brownlee 1996). This is quite remarkable because most chondritic meteorites do not typically have chondritic elemental composition at the 10 μm size scale. Most meteorites are coarser-grained and common 15 μm volumes are single mineral grains. Even the fine-grained matrix of carbonaceous chondrites is not “chondritic” in terms of elemental abundances. No mineral can approximate solar abundances for more than a few elements. The closest mineral match to chon-

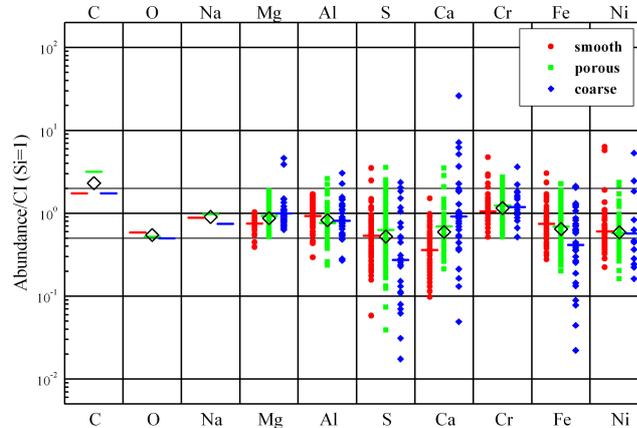


Figure 12. Bulk major element compositions of 200 (for C and O only 30) stratospheric IDPs showing a factor of two match with CI chondrites, some systematic offsets, and significant differences between smooth, porous, and coarse-grained particles. Horizontal bars for different sub-groups and diamonds for all particles are geometric means except for C, O, and Na, where only arithmetic averages were available. (Data from Schramm et al. 1989.)

dritic composition is $\text{Fe}_{50}\text{Mg}_{50}$ olivine ($\text{Fe}=\text{Si}=\text{Mg}$ atom fraction) that matches Fe, Mg, Si and O but severely misses for nearly all other elements. Typical IDPs are very fine-grained and even $15\ \mu\text{m}$ particles are assemblages of hundreds to millions of individual mineral grains plus amorphous material. This averaging effect in particles $> 2\ \mu\text{m}$ results in an approximately solar composition for the condensable elements. The elemental composition of typical IDPs is remarkably simple and is likely to be the result of a simple mechanical mixing of large numbers of randomly selected tiny grains.

1. Major Elements

The published data on IDPs include stratospheric particles as small as $5\ \mu\text{m}$ in diameter and particles as large as a millimeter collected from polar and deep sea sediments. The composition data have been obtained by a variety of techniques including the electron microprobe, X-ray fluorescence, neutron activation, and proton as well as synchrotron induced X-ray emission (PIXE and SXRF), and different types of secondary ion mass spectrometry (SIMS and TOF-SIMS). The largest data set for major elements (Schramm et al. 1989) comes from the EDX analyses of 200 stratospheric IDPs (Fig. 12). Brownlee (1997) gives electron microprobe data for 500 cosmic spherules in the $1\ \mu\text{m}$ to $1\ \text{mm}$ size range but the compositions of these melted particles only partly reflect that of their precursors because of effects related to hypervelocity entry into the atmosphere (Love and Brownlee 1991; Kornblum 1969).

The collected stratospheric IDPs appear to be a fair and probably the least biased sampling of the $15\ \mu\text{m}$ size meteoroids although some particles are excluded due to break-up by the stresses of atmospheric entry and col-

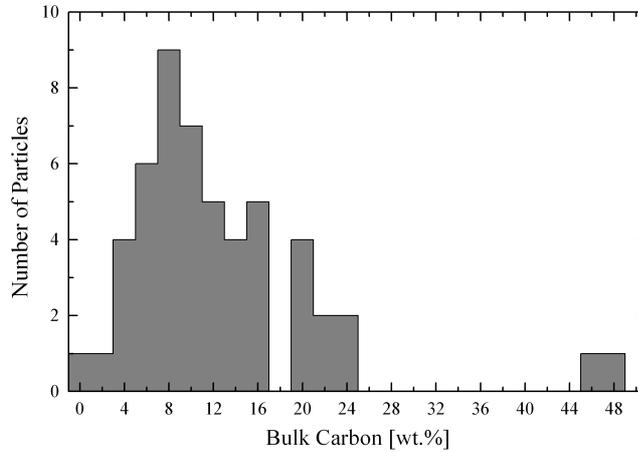


Figure 13. Bulk C concentrations of stratospheric IDPs (Keller et al. 1994). The peak of the distribution is more than twice the CI chondrites and the average is even higher.

lection. Particulate contamination problems are minimal during periods when the stratosphere is not loaded with volcanic particles. Any meteoroid type that would not be recognized or otherwise not be included in the collections probably represents less than 10 % of the 15 μm meteoroid population entering the Earth's atmosphere. There is evidence that some samples do, however, pick up contamination during their stratospheric residence (Jessberger et al. 1992; Bohsung et al. 1994, 1995a; Stephan et al. 1994a, b; Arndt et al. 1996a, b; Rost et al. 1996, 1999) but this appears to only severely affect trace elements, for example and most prominently Br (see below).

Studies of the major element abundances of 15 μm stratospheric IDPs show that most elements match the abundances of CI chondrites within a factor of two (Schramm et al. 1989). Analyses of large numbers of particles show peaking of compositions usually with only modest offsets from CI abundances. Some elements do, however, have abnormally large dispersions. Many particles show Ca and S depletions relative to CI abundances by a factor of two or more due to the high contents of phases such as FeS and carbonates in CI chondrites. Rare IDPs, however, have large *excesses* of these elements because they are largely composed of sulfide or Ca-rich minerals. If the data from many particles are combined, the depletions and excesses largely compensate for each other. C is an element that is commonly found at more than a factor of two above CI abundances (Fig. 13) (Thomas et al. 1993). C – in the sun after H, He, and O the fourth most abundant element – is highly fractionated among classes of chondrites, and it is significant that many IDPs have higher C abundances than the most C-rich meteorites. In fact, some IDPs are the most C-rich meteoritic materials and their high C abundance is strong evidence that they include samples of bodies so far not sampled by conventional meteorites.

2. *Minor and Trace Element Abundances*

Trace element data provide valuable information on IDPs. They may provide a means of defining sub-groups, indicate associations with chondritic groups, provide information on atmospheric heating and can act as sensitive tracers of the many potential processes that may have influenced IDPs from the time of formation to the time of sample analysis. While major elements are usually determined by standard electron probe techniques, analysis of minor and trace elements requires more elaborate methods such as PIXE (*Proton Induced X-ray Emission*), SXRF (*Synchrotron X-Ray Fluorescence*), and SIMS (*Secondary Ion Mass Spectrometry*), and this is more complex due to the more complex instrumentation and because of higher contamination hazards.

Recently a major compilation and investigation of trace elements, along with major and minor elements, from 89 IDPs has been published by Arndt et al. (1996a). They analyzed the abundances of 28 elements from Na to Zr, excluding the noble gases. The data for this enterprise mostly stem from three laboratories and – except some newly added data – had been published in the following papers: van der Stap et al. (1986); Wallenwein et al. (1987); Antz et al. (1987); Flynn and Sutton (1987, 1990, 1991, 1992a,b); Sutton and Flynn (1988); Stadermann (1990); Jessberger et al. (1992); Thomas et al. (1993, 1994); Flynn et al. (1993, 1994); Bohsung et al. (1995b); Arndt and Flynn (1995). The trace element data used by Arndt et al. (1996a) were obtained on stratospheric IDPs using PIXE, SXRF, and SIMS. Further techniques for IDP element analysis, not considered in this compilation, are: electron microprobe (Schramm et al. 1989; Thomas et al. 1993), time-of-flight (TOF) SIMS (Stephan et al. 1994a,b,c) and neutron activation (Zolensky et al. 1989; Lindstrom and Zolensky 1990). Unfortunately, a comparison of the combined data set with some of these techniques is problematic because of the very different depths (or volume) of information – many elements are rather heterogeneously distributed within the IDPs – and also because of the rather small overlap of the element list. On the other hand, a careful comparison of the PIXE, SXRF, and SIMS trace element results demonstrated (Arndt et al. 1996a) that these three techniques do not systematically differ, are coherent, and thus can be compared and combined to a single full *complete* data set.

In order to allow this comparison and the further investigation of that data set, the published material was recalculated to one common dimension: element abundance normalized to Fe and CI chondritic abundance taken from Anders and Grevesse (1989). In cosmochemistry generally Si is taken for normalization. However, Si data were published only for about 50 % of the IDPs, the quality of Si data obtained by PIXE and SXRF is rather poor due to absorption effects, and – a dilemma for all analytical techniques – because the probable contamination with silicone oil from collecting and maneuvering the IDPs.

The results of the study are shown in Fig. 14. On the average S, Ca, and Ni are depleted in IDPs relative to CI chondrites while the eleven elements

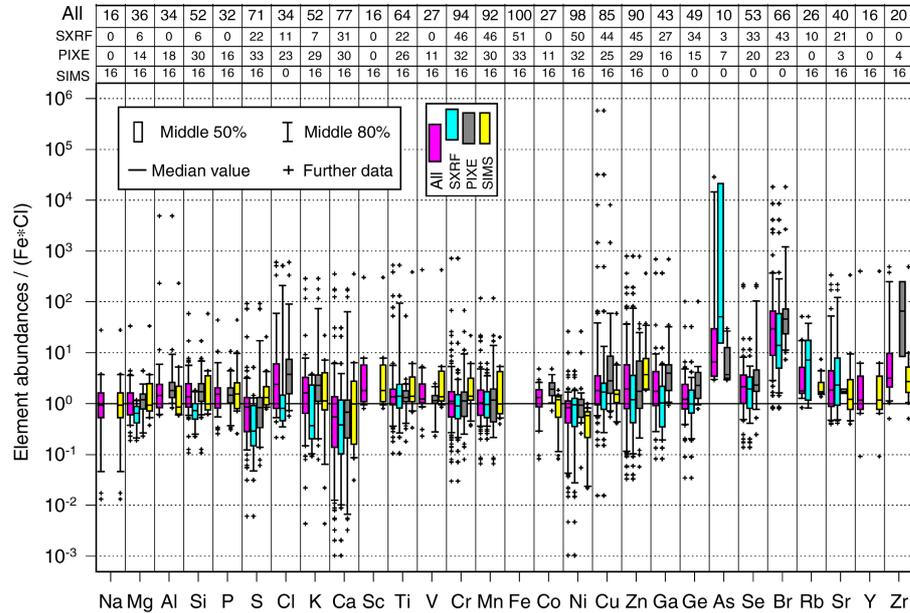


Figure 14. Element abundances of 100 bulk analyses of 89 IDPs normalized to Fe and the respective CI chondritic element/Fe ratios. The input for this plot constitutes *all* presently available major, minor, and trace element abundances in IDPs. The data set is broken down according to the three employed experimental techniques: PIXE, SXRF, and SIMS. The data are presented in a boxplot where “further data” indicates those that are outside the 80 % range. On the top line the number of available entries for the different methods are given. (From Arndt et al. 1996.)

from Cu to Zr are enriched. The abundances of the remaining elements are indistinguishable from that of CI-chondrites and thus from the mean solar system. Arndt et al. (1996a) – in order to check for bimodal or multimodal patterns – examined the distributions of the element abundances and found only unimodal distributions with the exception of Ni that has a peak near the CI-ratio and a small group with low Ni values.

The data set was also analyzed with cluster analytical techniques to search for groups and with discriminant analysis to investigate their separation (Masart and Kaufmann 1983; SAS Institute Inc. 1988). The aim of cluster analysis is to find in multidimensional space groupings within a given data set that are (a) as homogeneous as possible, (b) separated from each other as far as possible, and (c) based on as many variables as possible. The main problem in the cluster analysis of the IDP data set stems from missing data: Only *one* single element, Fe, was measured in *all* particles. In fact, 50 % of the possible element data table is empty because of insufficient sensitivities of the instruments and abundances below detection limits. As a consequence, (a) no cluster analysis could include all IDPs simultaneously, (b) what is known about the cosmochemical nature of the elements had to be ignored and (c) certain ele-

ment combinations had to be chosen to encompass the maximum number of IDPs.

A cluster analysis based on the six elements Cr, Mn, Ni, Cu, Zn, and normalizing Fe, cluster 82 % of the IDPs into four groups which were named chondritic, low-Zn, low-Ni, and nonsystematic. The geometric mean abundances of all 28 elements within three groups – chondritic, low-Zn, and low-Ni – are given in Fig. 15. In the left hand panel of Fig. 15, the elements are plotted against their volatility taking as its measure the 50 % condensation temperature in a gas of solar composition as compiled by Wasson (1985). In the right hand panel, the elements are plotted against decreasing CI abundance. In order to avoid over-interpretation of less significant mean abundances, a light gray symbol is used if an element is detected only in less than half of the IDPs in the group, otherwise the symbol is black. In the following we discuss the four groups.

Chondritic Group

The abundances of many elements in the chondritic group (44 members) are rather close to CI but with frequent and major depletions of Ca and S ($0.8\times$ CI). Four elements are very close to CI: Mg, Mn, Ni, and Y. Enrichments are found at three levels: (a) most prominently ($5\text{--}30\times$ CI) for Br, As (only rarely detected), and Zr, (b) in the narrow range $2.2\text{--}2.7\times$ CI for the elements Cl, Cu, Zn, Ga, Se, and Rb, and (c) in the equally narrow range $1.2\text{--}1.7\times$ CI for Na, Al, Si, P, K, Sc, Ti, V, Cr, Co, Ge, and Sr. Some of the Si overabundance, however, probably stems from the unavoidable silicone oil. The fact that the latter large number of elements is tightly grouped around $1.4\times$ CI may result from the Fe-normalization and may indicate an overall Fe-deficiency of $0.7\times$ CI. Such a deficiency is supported by the few available absolute Fe concentrations ($0.9\times$ CI). If the Fe-deficiency of $0.7\times$ CI is indeed true then (a) 16 elements, (Na–Sr) and (Mg–Y), in IDPs are present in solar abundance within 30 %, (b) Ca and S deficits increase, and (c) the enrichments of the second group of elements (Cl–Rb) decrease to $1.7\times$ CI. It has clearly been demonstrated (Arndt et al. 1996a) that at least PIXE is sufficiently sensitive to detect such enrichments of Cl, Cu, Zn and Se, possibly also Ga, and marginally Rb. The elements enriched in this group are also most strongly enriched in the group of the non-chondritic particles.

Low-Zn Group

Ten of the twelve particles in that group show features that previously had been interpreted as depletions due to atmospheric entry heating. The elements detected in at least six out of the twelve IDPs can be ordered in terms of their abundance relative to Fe and CI chondrites: highly enriched is again Br; almost normal are Ga, Se and Cu; highly depleted are Ca and Zn; moderately depleted are S, Cr, Mn, Si, Ti, and Ni ($0.26\text{--}0.61\times$ CI). Assuming that Fe is enriched by a factor of 2.5 in the low-Zn group compared to the chondritic group, an excellent agreement between these two groups is found, except a

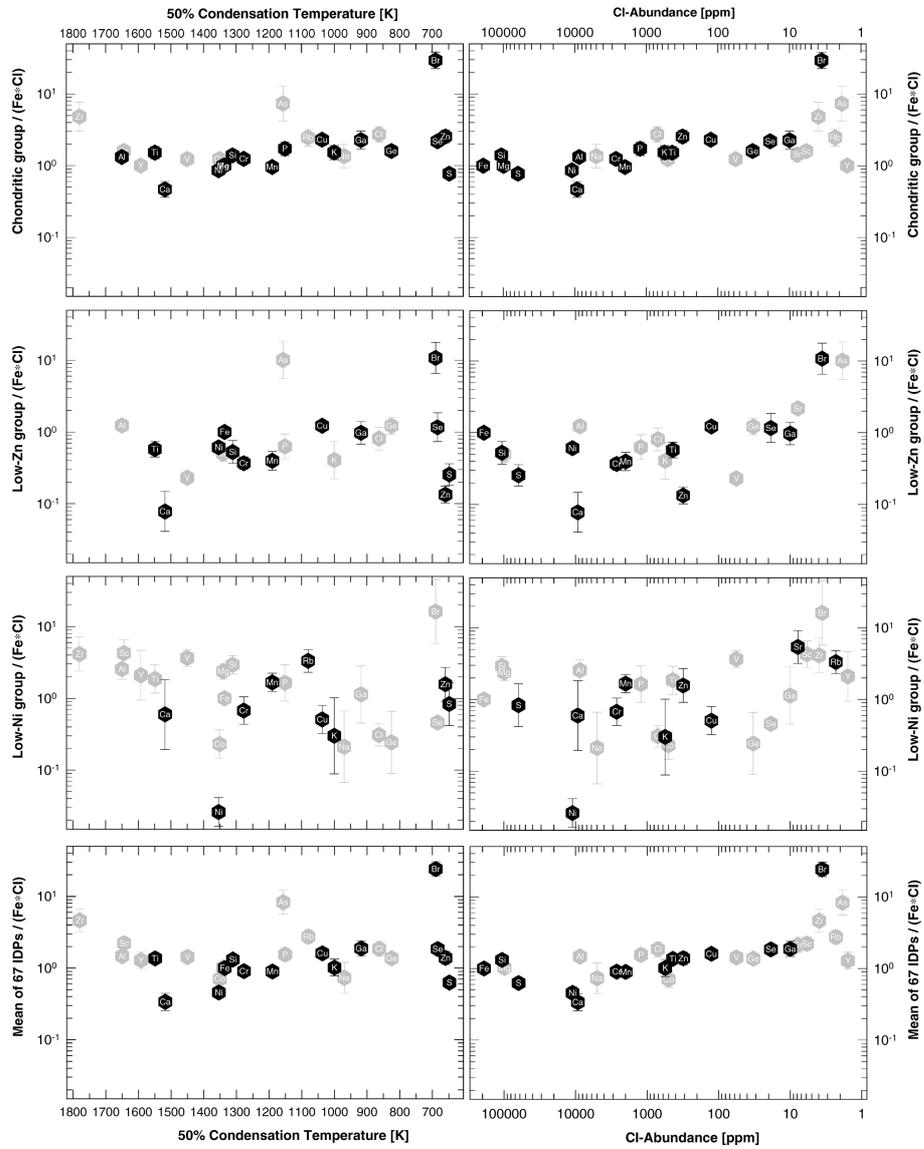


Figure 15. Geometric means and standard deviations of the element ratios within the chondritic, low-Zn, and low-Ni IDP groups (from top). The bottom panel shows the geometric means of IDPs of these three groups. In the left column the abscissa is the 50 % condensation temperature (Wasson, 1985) and in the right column the respective CI-chondrite abundance (Anders and Grevesse 1989), both decreasing to the right. Symbols of elements that are detected in < 50 % of the IDPs of the respective group are gray, the others black. (From Arndt et al. 1996a.)

probably insignificant difference in Ca. The only really significant difference between the two groups is the abundance of the volatile element Zn. Most importantly, there is no difference in two other volatile elements that have about the same condensation temperature as Zn: S is depleted in both groups by the same factor and Se in both groups is present in nearly CI abundance.

Low-Ni Group

The mean value of Ni/Fe in the low-Ni-group (11 members) is $0.03 \times \text{CI}$. This group is quite variable with considerable spread in the element means. Some member IDPs of this group are discussed by Flynn and Sutton (1990). They describe element patterns remarkably similar to that of basalt and unless solid proof of extraterrestrial origin is available, e.g., solar flare tracks or solar wind He, some of the low-Ni particles could be terrestrial. In these particles Ca/Fe is higher than in the previous two groups and overlap the CI-value. High mean values for a number of elements indicate a generally low Fe content. This is supported by the Fe mean ($0.7 \times \text{CI}$) based on the (few) absolute Fe concentrations. Three particles have low Zn *and* low Ni values.

Non-Systematic Group

This is the most weakly defined group with only six members that are most distinct from all the other particles. The group is characterized by enormous enrichments of the same elements that are also heavily enriched in the other groups: Br, Al, Cu, Zn, Ga, Se, Cl, and Si (in decreasing order).

In conclusion, the cluster analysis based on Cr, Mn, Ni, Cu, and Zn reveal four groups of stratospheric particles, showing some common but also some distinguishing features: The fourteen most abundant elements (Fe, Si, Mg, S, Ni, Ca, Al, Na, Cr, Mn, P, Cl, K, Co) in chondrite-like IDPs are as abundant as in CI chondrites and the solar system in general, only Ca and S are depleted. This does not hold for the low-Ni particles with Ni depletions and about normal Ca, and for the low-Zn group with high Fe. The remaining fourteen trace elements (Ti, Zn, Cu, V, Ge, Se, Ga, Sr, Sc, Zr, Br, Rb, As, Y) in *all* three groups are enriched by about the *same* factor, $2 \times \text{CI}$, with the exception of Br with enormous enrichments of typically about $25 \times \text{CI}$ but up to $20\,000 \times \text{CI}$.

The fourteen elements that on the average in chondrite-like IDPs are present in cosmic abundance set the baseline for enrichments of the many and depletions of the few elements. Deviations from the solar abundances may be indigenous and may be related to the origin and history of IDPs, but almost certainly they are at least in part the result of processes in the Earth's atmosphere and even on the Earth. In addition, the magnitudes of element losses and gains certainly depend, besides pre-atmospheric size and density, on entry velocity and angle, but also on the chemical compound the elements are in, all of which is unknown for individual IDPs. Some contaminating processes, however, can be studied as will be demonstrated next.

3. Contamination of IDPs

The research on the IDP contamination problem mostly rests on PIXE and TOF-SIMS analyses of minute sub-volumes of IDPs. Since TOF-SIMS, a technique primarily developed for surface analysis (Benninghoven 1994), is a rather novel system it will shortly be introduced. A flock of 500 primary ions, often Ga^+ , hit with a repetition frequency of 5–10 kHz an area with $\sim 0.2 \mu\text{m}$ diameter of the surface to be studied. About ten secondary ions per shot are released and, after mass separation in a drift tube equipped with a reflector, are detected with a time resolution of typically 200 ps. That together with the primary ion pulse length of ~ 1 ns results in a mass resolution of several thousand. Spectra and secondary ion images – of positive as well as negative polarity in subsequent analyses – are generated by integrating minutes to hours of mass separated secondary ion counts. The high spatial resolution provides images of the lateral element distribution as well as quantitative measurements of element ratios on sections and surfaces of individual IDPs. Simultaneous detection of all secondary ions, sequentially with both polarities, nearly infinite mass range, an instrumental transmission of 20–80 %, and the small sample consumption that can be restricted to monolayers ensures that a maximum chemical information can be obtained without completely destroying even small samples. The lateral distribution of elements and molecules is investigated by scanning the samples with the primary ion beam (Schwieters et al. 1991). Being a surface analysis technique, the information depth is in the order of a few monolayers. Thus, TOF-SIMS is an important technique to study, among others, surface features like contamination of IDPs.

Contamination by silicone oil presents an obvious problem for *all* stratospheric IDPs since they are captured in silicone oil and are also immersed in it for manipulation.

Another most prominently enriched element in IDPs is Br. This has been noted in very early IDP trace element analyses and then has been interpreted to indicate large scale chemical heterogeneity of the early solar system (van der Stap 1986). Later, however, this contention was disputed based on indications that atmospheric contamination processes played an – up to that time underestimated – role in the trace element budget of stratospheric IDPs (Jessberger et al. 1992). From then on evidence is growing that at least the Br overabundance is due to halogen-rich aerosols although the halogen abundance in the stratosphere is not very well known (Cicerone 1981). First direct observational hints for such contamination of chondritic IDPs were Br-salt nanocrystals attached to IDP W7029E5 (Rietmeijer 1993). Stephan et al. (1994b) demonstrated by TOF-SIMS studies of a section of IDP L2006G1 that it has a halogen-rich exterior rim. The TOF-SIMS images (Plate 2) of most ions resemble a ring structure with C and H in the center. Outside the ring, Cl and F are concentrated on one side of the rim of the particle. In this rim Br is clearly identified. In a further study (Stephan et al. 1995), a set of ten stratospheric particles was selected first for bulk SEM-EDX including C measurements (Thomas

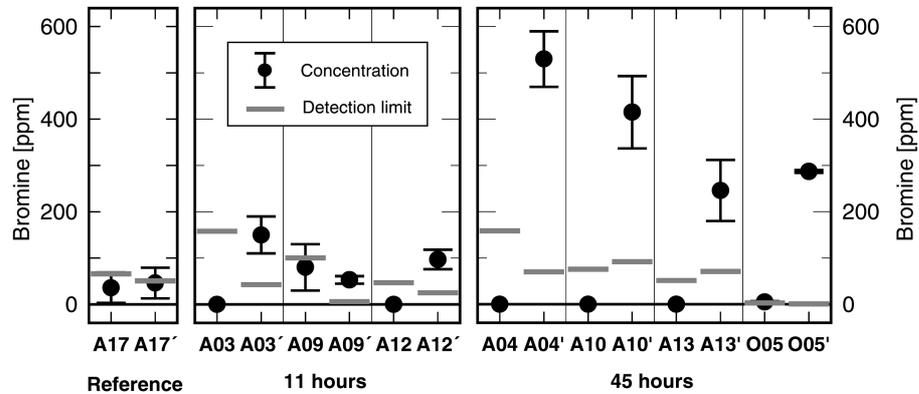


Figure 16. Br concentrations, obtained with PIXE, in seven aerogel (A##) and one Orgueil (O05) particles before (plain identification number) and after (number with ') stratospheric IDP collection flight. The limits of detection, LODs, are also indicated. They depend, among other factors, on the total proton dose and thus are variable. The duration of stratospheric exposure is given at the bottom. Aerogel particle A17 was not flown but was subject to all other transport and handling procedures. Br contamination occurred during stratospheric flight, and its magnitude is related to the flight duration. No contamination by transport and handling is observed.

et al. 1994) and then for TOF-SIMS analyses of the *very* surfaces of these IDPs (Rost et al. 1996, 1999). With such surface TOF-SIMS analyses, because of its roughness a challenge for the method, clearly the presence of F, Cl, and Br on IDP surfaces was indeed and without doubt demonstrated (Plate 3). Except for U2015G1 where Br was found in an extremely fine-grained iron oxide sub-unit (Stephan et al. 1994a), none of the particles analyzed so far with TOF-SIMS showed appreciable amounts of Br in the bulk.

Arndt et al. (1996b, 1997) imitated systematically that portion of the IDP history that can be imitated: the time on the IDP collector and the laboratory handling up the analysis. They exposed various small (10–40 μm) and porous particles – among them aerogel and CI chondrite fragments – to the stratosphere for 11 and 45 hours on an actual IDP capture flag. All particles were analyzed prior to flight with PIXE, some particles were just handled, but kept on the ground for a laboratory contamination test, and the recovered particles were again analyzed with PIXE thereafter. Br that was not detected before flight was clearly present in all particles after exposure to the atmosphere (Fig. 16). PIXE-mapping of one of these particles (Fig. 17) demonstrates that Br is exclusively confined to the particle and is absent in the surrounding material, also absent in silicone oil. This finding again corroborates the earlier contention (Jessberger et al. 1992) that contamination processes indeed play an important role although the detailed mechanisms are not yet fully understood.

Certainly, the effectivity of contaminating processes depends on a number of factors like chemical composition and state or on porosity and size of the IDP constituents (Stephan et al. 1994a) that largely remain unknown at present.

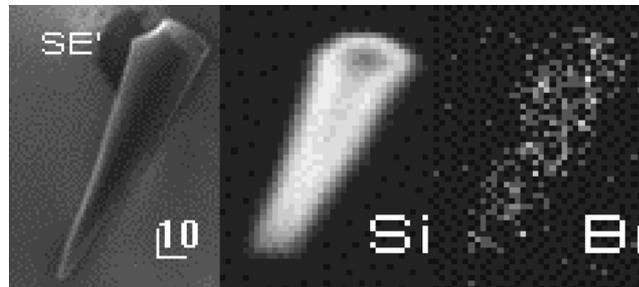


Figure 17. Secondary electron image (left) and Br distribution (right), both obtained with PIXE, of an aerogel particle and its surrounding after 45 hours of stratospheric collection flight. The scale bar length is $10\ \mu\text{m}$. The beam size for the Br-mapping was several microns. Br is exclusively found in the aerogel particle.

In addition, it must be seen what other trace elements that are also enriched in IDPs – though to a much lesser degree than Br – are affected as well (Arndt et al. 1997). Therefore, to obtain definitive information on the siting of the enriched as well as the depleted elements is a crucial goal of future IDP research that can, however, be achieved by compositional mappings with PIXE and TOF-SIMS in addition to other microanalytical techniques like ATEM.

4. Compositional Mapping

PIXE at the Heidelberg proton microprobe allows elemental mapping with a lateral resolution down to $1\ \mu\text{m}$ and thus provides additional information on the siting and distribution of the elements without destroying the particle before or during the analysis (Maetz et al. 1996). An example of a Zn and Fe-rich IDP is given in Fig. 18. Here, Fe is very inhomogeneously distributed and correlated to Ni and S, but anti-correlated to Mg and Si. The Zn-enrichment ($10\times\text{CI}$) is exclusively confined to the FeS-Ni region in the center, a possible siting for Zn in IDPs (Arndt et al. 1996a).

From imaging TOF-SIMS analyses a few element maps are presented here. Plate 4 shows secondary ion images for 18 different ion species from particle L2006E10. Here the correlation of Ca and P, both highly enriched in a $4\times 1\ \mu\text{m}^2$ sized region, lead to the discovery of presumably an apatite grain in a hydrated IDP (Stephan et al. 1994b).

The rather unique and long list of elements accessible with TOF-SIMS – from the light volatile to the heavy refractory elements – allows a comparison with the composition of comet Halley's particles (Jessberger et al. 1988, 1989). The composition of the only anhydrous IDP analyzed so far with TOF-SIMS (L2006B21; Stephan et al. 1993) and Halley's dust data – as far as they are available – agree relatively well at least for some elements (Fig. 19). The general trend for Halley's element/Si ratios normalized to CI chondrites, characterized by C-enrichment and depletion of elements like O, Mg, Al, Cr, Mn, Fe, and Ni, as well as almost chondritic Ti and Co, was also found for L2006B21, although to different extents. Future analyses with TOF-SIMS of anhydrous IDPs may

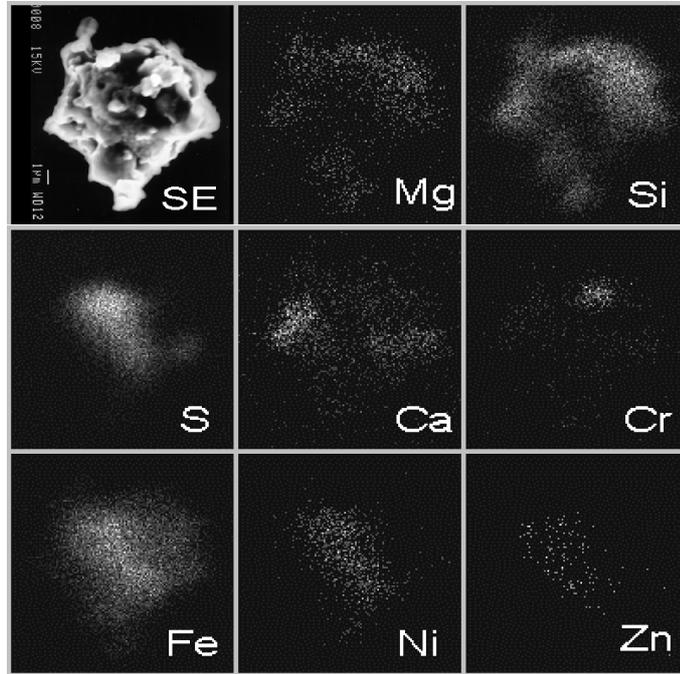


Figure 18. SEM-image (top left; W. Klöck pers. communication) and element mappings obtained by PIXE of eight elements in IDP L2005 AD14. The proton beam size was $1.5 \mu\text{m}$, the pixel distance is $0.1 \mu\text{m}$. Note the interior bar rich in Fe, S, Ni, and Zn.

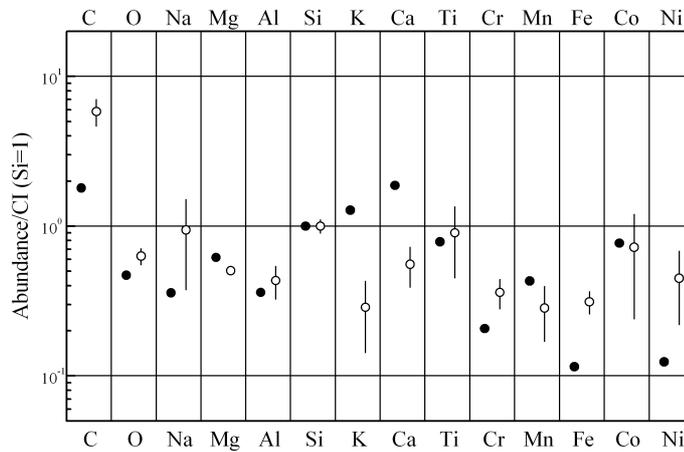


Figure 19. Comparison of element ratios – normalized to CI-chondrites and Si for anhydrous IDP L2006B21 (obtained with TOF-SIMS; filled circles) and comet Halley's dust (open circles). It should be noted that the cometary composition is uncertain by about a factor of two as discussed in detail by Jessberger et al. (1991); cf. also Sekanina et al. (1998), and that the indicated error bars reflect the variability of the cometary dust compositions.

help to solve the question whether anhydrous IDPs originate from comets, provided that more and more secure data on the chemical composition of different comets will be available. Here TOF-SIMS also may play an important role on future space missions for the *in situ* analysis of cometary matter (Beck and Kissel 1994).

III.F. Isotopic Composition

Studying the isotopic compositions of IDPs is of particular interest for two main reasons: First, isotopic signatures can prove that a given particle is actually extraterrestrial. Since the isotopic compositions of the elements in natural and most artificial terrestrial dust particles vary only within small margins – variations from radioactive decay are certainly not considered here – any composition clearly outside that range (i.e., an isotopic anomaly) represents proof of a particle's extraterrestrial origin. However, the reverse is not true and the absence of isotopic anomalies in a given particle can not establish its terrestrial origin. The possibility to positively prove the extraterrestrial origin of particles on an individual basis is particularly important because all currently available collection sites, including the stratosphere and the Low Earth Orbit (LEO), are subject to natural terrestrial and man-made contamination. The second reason for performing isotopic measurements in IDPs is to gain information on the origin and history of a particle. When the physical and chemical processes that lead to an observed isotopic anomaly are known, elements of the history of an IDP and/or its progenitor material can be inferred.

For bulk samples with macroscopic dimensions a number of mass spectrometric analytical techniques exist that allow routine precision measurements of isotopic compositions. However, when it comes to the analysis of particles with micrometer-dimensions, only a few methods offer the required sensitivity. Fortunately, many of the isotopic anomalies observed in IDPs are so pronounced that a slightly lower measurement precision than for the analysis of terrestrial samples can be accepted. The first isotopic measurements of IDPs were made with thermal ionization mass spectrometry (Esat et al. 1979; Esat and Taylor 1987). More recently, step-heating and pulse-heating techniques have been used to determine the noble gas contents of IDPs, and to some extent their noble gas isotopic compositions (Nier 1994). However, the vast majority of all isotopic measurements of IDPs has been made with double-focusing secondary ion mass spectrometry (SIMS) due to its unique ability to measure isotopes at low abundances with high spatial resolution and sufficient precision.

The distinction of whether or not an observed isotopic composition is anomalous requires the knowledge of the range of compositions that is considered *normal*. With respect to extraterrestrial materials, any isotopic composition found in terrestrial samples is seen as *normal*, i.e., the terrestrial samples with the most diverging isotopic compositions define the normal terrestrial range for that particular element. Detailed information on terrestrial isotopic ranges for a number of elements can be found in e.g., Hoefs (1980). An isotopic composition is considered anomalous, when it is separated by more than two

sigma from the normal terrestrial range (McKeegan 1987a). This definition of an isotopic anomaly is distinctly different from what is generally being used in terrestrial geochemistry.

An isotopic composition is frequently expressed as a delta-value, which is the deviation of the measured isotopic ratio from a standard ratio in per mill units. Isotopic ratios are generally calculated with respect to the most abundant isotope in terrestrial standards. Thus, e.g., the expression $\delta^{13}\text{C} = 200\text{‰}$ indicates that in this sample the $^{13}\text{C}/^{12}\text{C}$ ratio is 200 parts per thousand (= 20 %) higher than the standard $^{13}\text{C}/^{12}\text{C}$ ratio. For many elements a standard isotopic ratio with a δ -value $\equiv 0\text{‰}$ has been defined, e.g., SMOW (standard mean ocean water) for H and O (Hagemann et al. 1970), PDB (Pee-Dee Belemnite) for C, and air for N. The results of isotopic measurements in IDPs are listed in the following paragraphs by element.

Hydrogen is particularly susceptible to isotopic fractionation because of the large relative mass difference of the isotopes D and H. This also increases instrumental mass fractionation effects, making the measurement of H isotopic compositions less precise than those of other elements. Zinner and McKeegan (Zinner et al. 1983; McKeegan et al. 1985, 1987) measured the H isotopic composition in 31 chondritic IDPs with SIMS. In 13 particles anomalous D/H ratios with $\delta\text{D} > 100\text{‰}$ and in three particles with $\delta\text{D} > 2\,000\text{‰}$ were determined. In five particles it could be demonstrated that D-excesses are spatially correlated with the C abundance, hinting at a carbonaceous carrier of the anomaly. Only IDPs of the hydrated silicate variety and the *pyroxene* type were found to exhibit H isotopic anomalies, but not *olivine* type IDPs. In a follow-up study, covering some of the same particles and a number of new ones, Stadermann (1990) also observed D anomalies in roughly 1/3 of the analyzed chondritic IDPs. The largest D/H-ratio found in this study was $\delta\text{D} = 1\,100\text{‰}$. The results of the H measurements indicate that many IDPs are isotopically very heterogeneous, with some fragments having essentially normal D/H-ratios while other fragments of the same particles were enriched in D with $\delta\text{D} > 1\,000\text{‰}$. An important discovery was made by Messenger et al. (1996) when he observed that H isotopic anomalies were much larger and more common in so-called "cluster IDPs" than in individual, non-cluster IDPs. These cluster IDPs are particles that fragment during impact on the stratospheric collector flags. While it is not obvious why cluster particles *per se* should exhibit larger anomalies than other IDPs, there is a hint towards a time-dependence in the data set, which could indicate a correlation with a specific dust source (Messenger and Walker 1997).

The large variations of the H isotopic composition made it interesting to study these samples with SIMS isotope imaging. In this technique, secondary ion images of the distribution of different isotopes on the sample surface, such as D and H in this case, are mathematically converted point-by-point into a distribution image of the isotopic ratios on the surface of the sample. These images can then be presented as false-color pictures, directly showing the isotopic heterogeneity of the sample. Measurements of this kind have first been made

by McKeegan et al. (1987) in a chondritic IDP called Butterfly. This particle showed a local D enrichment (hotspot) with $\delta D > 9\,000\text{‰}$ concentrated in a spot with a diameter of $1\ \mu\text{m}$ or less. Similar imaging measurements have also been made by Fleming et al. (1989) in IDPs with smaller, but significant H isotopic heterogeneity. Recent H imaging studies by Messenger (Messenger et al. 1996; Messenger and Walker 1997) identified D-hotspots with $\delta D > 50\,000\text{‰}$. Those are the largest D excesses ever observed in a natural sample.

Lithium is still light enough to make SIMS measurements relatively easy due to the fact that there are only few interferences at the atomic masses 6 and 7. Stadermann (1990) measured the ${}^6\text{Li}/{}^7\text{Li}$ -ratios in 12 stratospheric dust particles without finding any compositions significantly different from that of the terrestrial standard. Xu et al. (1994) determined the ${}^6\text{Li}/{}^7\text{Li}$ -ratios in four IDPs and found general agreement with the values in reference literature. Boron was also analyzed in that study by Xu et al. (1994) who measured the ${}^{10}\text{B}/{}^{11}\text{B}$ -ratios of the same four IDPs and again found agreement with reference data.

Carbon that is found to be isotopically anomalous in presolar grains extracted from certain meteorites (Anders and Zinner 1993) was first studied in IDPs by McKeegan (1987a). He determined the ${}^{13}\text{C}/{}^{12}\text{C}$ -ratios in seven IDPs and found an anomalous composition in only one case, in fine-grained chondritic material adhering to a refractory Al-oxide particle. In 124 measurements on fragments of 68 stratospheric dust particles Stadermann (1990) did not observe any C isotopic anomalies, although several of the same fragments showed large ${}^{15}\text{N}$ excesses in the same measurements. With respect to their C isotopic composition, Messenger (1997) found that cluster IDPs are normal and indistinguishable from individual IDPs.

Nitrogen isotopic compositions of IDPs were first measured by Stadermann (1990). He found significant ${}^{15}\text{N}$ enrichments with $\delta^{15}\text{N}$ values up to 442‰ in 12 out of 35 chondritic IDPs as well as in one overall chondritic particle that, however, was significantly enriched in Na, K, and P. Due to the nature of these SIMS measurements, where N is detected as CN^- ions, the observed anomalies only represent N in association with a C-rich phase. Messenger (1997) also observed heavy N in several IDPs with ${}^{15}\text{N}$ enrichments up to $\delta^{15}\text{N} \approx 500\text{‰}$. Most analyzed IDPs appear to be isotopically very heterogeneous with respect to N, similar to what was observed for H. A correlation between the isotopic anomalies of H and N, though not in their magnitude, was found by Stadermann (1990). Most particles with N anomalies also have H anomalies but the inverse is not true (Messenger et al. 1996). In cluster IDPs $\delta^{15}\text{N}$ enrichments were found to be more common than in non-cluster IDPs, but the effects were not generally larger (Messenger 1997).

Oxygen isotopic anomalies with substantial ${}^{16}\text{O}$ enrichments have been reported for a few IDPs with a refractory elemental composition. The extend of the anomalies and the composition of these IDPs strongly resemble refractory meteorite inclusions (McKeegan 1987b; Stadermann 1990). Significant O iso-

topic anomalies have not been observed in any chondritic IDPs, however, only very few particles have been analyzed (cf. Greshake et al. 1996).

Magnesium and *Silicon* isotopic compositions were measured in several particles by McKeegan (1987a) and Stadermann (1990). No isotopic anomalies besides some terrestrial, i.e., mass-dependent fractionation were found. Evidence of ^{26}Mg excesses at the per mill level has been reported earlier by Esat et al. (1979).

The major isotopic anomalies found in IDPs are those of H, N, and O. For both H and N, ion-molecule exchange reactions in a cold interstellar molecular cloud can explain the observed isotopic compositions (Zinner 1988; Tielens 1997). However, the same type of reactions are also expected to lead to isotopic fractionation of C that, however, has not been detected in IDPs, even though the C isotopic composition of a large number of particles has already been measured. One possible explanation for the lack of C isotopic anomalies is that C could be fractionated through several chemical routes with opposite and partially canceling effects (Tielens 1997). The evidence for molecular cloud material in IDPs and meteorites has recently been reviewed by Messenger and Walker (1997). The observation of deuterium hotspots with enrichments up to 50000 ‰ in IDPs with the ion imaging technique has prompted similar studies in primitive meteorites. Guan (1998) indeed found comparable hotspots with δD up to 8000 ‰ in black fragments in the matrix of the CR2 chondrite Renazzo. So far, the D carriers in meteorites and IDPs are not positively identified.

Future isotope studies of IDPs should attempt to answer the following questions: Can isotopic anomalies of other elements be found in IDPs, e.g., by extending the precision of the measurements or by extending the number of elements measured? Which are the carrier phases of the anomalies? What are the exact mechanisms leading to observed anomalies and why do certain elements show normal isotopic compositions while others in the same IDP are anomalous? At least with respect to the first two questions, improvements in micro-analytical measurement techniques will be beneficial.

IV. ORIGINS

For distinguishing different particle classes that might have genetic implications, it does not appear that major element composition is a strong discriminator. Small samples of primitive solar system materials, to first order, tend to have similar compositions with scatter due to size-scale dependent heterogeneities in their parent bodies. Thomas et al. (1995) have shown that considerable insight into these heterogeneities within IDPs can be obtained by the analysis of so-called “cluster” particles that fragment into large numbers of $> 5 \mu\text{m}$ components during collection in the stratosphere. Compositional ranges in these particles are similar to the overall range of individually collected particles. The natural level of compositional variation of nanogram particles produce a “noise” background that complicates accurate compari-

son with meteorite groups where characteristic compositions are derived from much larger samples. Physical properties such as structure and mineralogy do provide strong clues to origin and evolutionary processing (Klöck and Stadermann 1994) but at least to first order, major element abundances do not seem to be highly diagnostic properties. This is in contrast to meteorites where elemental composition is an important criterion for classification.

The major sources of these particles most probably are asteroids and comets (Leinert and Grün 1990). Especially the latter makes IDPs exciting since they provide us with the opportunity to study cometary matter in the laboratory and hence in very much greater detail than, e.g., in previous space missions to comet Halley (Jessberger et al. 1988; Jessberger and Kissel 1991). Cometary IDPs eventually may provide information on matter from (a) the ancient solar system, (b) the outer solar system and (c) even from pre-solar epochs (Brownlee 1994). But before such information can be gathered, the fundamental, yet unsolved question in IDP research must be answered: Which IDPs are cometary and which are asteroidal?

One criterion to distinguish particles from the two sources is the velocity at atmospheric entry. Model calculations demand that cometary particles in general have higher entry velocities than asteroidal particles, and that there is only a small velocity overlap between both groups (Jackson and Zook 1992). Therefore, cometary particles are expected to be more strongly heated upon atmospheric aero-braking than asteroidal particles (Flynn 1989). Consequently, one aim in IDP research is to determine the degree of heating a particle has experienced. Maximum temperatures can be estimated if the particles still contain solar flare tracks (Bradley et al. 1984). A minimum entry temperature indicator may be the degree of loss of volatile elements. Such losses are observed for the volatile elements S, Se, Ge, and Zn in heating experiments with 100 μm fragments from the CI chondrite Orgueil taken as analogues of Antarctic micrometeorites (Greshake et al. 1998). Zn abundances in IDPs that are significantly lower than in CI chondrites have been interpreted as indicating Zn losses and thus were related to high entry temperatures (Flynn and Sutton 1992a). Also discussed by some researchers is the presence of magnetite rims as implying strong entry heating (Klöck et al. 1992) but alternative interpretations of these rims are advocated by others (Maurette et al. 1993). A possibly powerful technique to estimate minimum temperatures is provided by He stepwise degassing experiments (Nier and Schlutter 1993) that provide diffusion information – and thus an estimate on the duration and temperature at atmospheric entry. It turns out that if solar wind ^4He is lost, an IDP must have experienced typically at least 700°C. At such a high temperature, however, particles are conjecturally rather strongly modified especially as far as volatile or organic components are concerned. As judged at first glance from the Halley results (Jessberger et al. 1988; Krueger et al. 1991) these components probably are the ones that are most different in comets compared to meteorites. Consequently, the He method rejects IDPs as being unaltered cometary matter

and does not help to identify pristine cometary particles. This applies to all methods using temperature criteria.

Then the question arises if there are chemical groups that are significantly special and that eventually may be related to cometary and asteroidal sources, respectively. One might expect that the more elements are included in tackling that problem the higher the chances are to approach an answer. A few research groups are engaged since the mid-1980s to obtain trace element concentrations in IDPs (Arndt et al. 1996a), but hitherto without plain evidence as has been detailed above.

All in all, the years of effort of quite a number of research groups did not yet result in establishing clear-cut criteria as to the source of any individual IDP or of certain groups of IDPs. Surely, one rather technical reason for this shortcoming is the tiny dimension of IDPs that makes concerted analyses with complementary methods pretty difficult. Thus they are very rare. One may expect, however, that the presently ever accelerating technologic advancement in the various required micro-analytical methods (Jessberger 1991; Walker 1991) eventually leads to better substantiated and more complete known facts on individual IDPs. A second reason is very obvious: We do not know enough about comets. Unquestionably, the experiments on the missions to comet Halley provided a wealth of unprecedented data. However, the inherent uncertainties of the elemental abundances of Halley's dust, the vague information on the molecular inventory, the very scarce knowledge of the isotopic composition (except a few high $^{12}\text{C}/^{13}\text{C}$ particles; Jessberger and Kissel 1991), and – very important – the absolute lack of structural and mineralogical information obstructs comparison with facts obtained on IDPs in sophisticated laboratories on Earth. Thus, the question of the source of an individual IDP is not answered at present, and probably must wait for the results from future space missions to comets that also include retrieval of cometary dust.

Acknowledgements

This work was supported by FWF and ÖAD in Austria, and by the Deutsche Forschungsgemeinschaft as well as by the Ministerium für Schule und Weiterbildung, Wissenschaft und Forschung des Landes Nordrhein-Westfalen in Germany.

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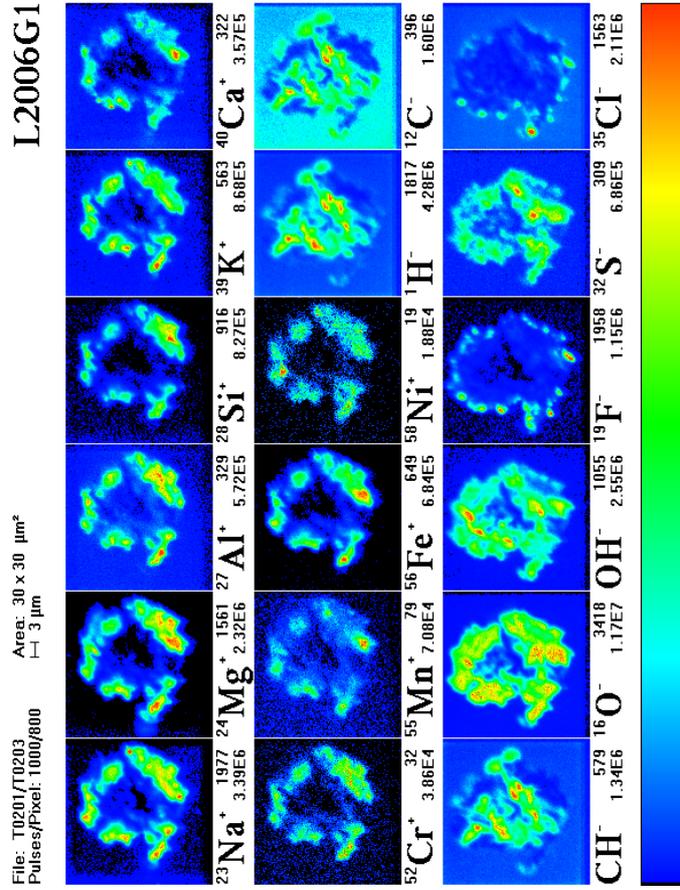


Plate 2. (Chapter on Properties of Interplanetary Dust by Jessberger et al., p. 278). TOF-SIMS secondary ion images for the IDP L2006G1. Positive and negative ions were measured consecutively. The field of view ($30 \times 30 \mu\text{m}^2$) and the number of primary ion pulses per pixel (1000 for positive, 800 for negative secondary ions) are given on top. The sample was scanned (128/128 pixels) with a Ga^+ primary ion beam. The subscripts of the individual ion images give the secondary ion species (e.g., $^{23}\text{Na}^+$), the maximum number of counts/pixel (1977) and the integrated number of counts for the total image ($3.39 \cdot 10^6$). For all secondary ion images the same linear color scale was used, where black corresponds to zero counts and the maximum count rate is red.

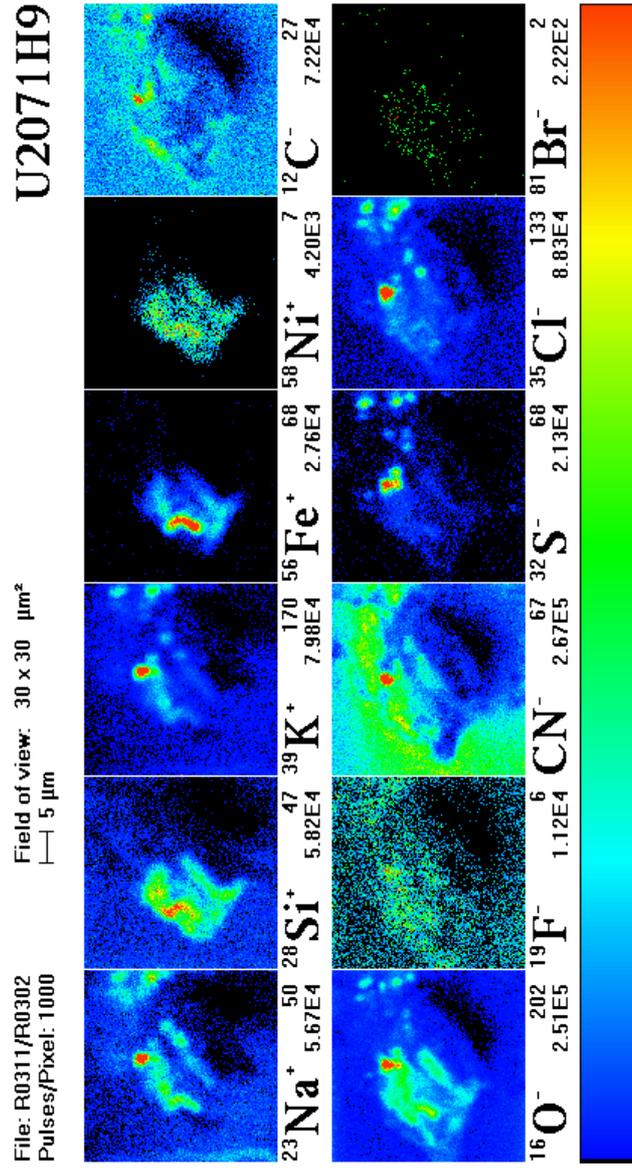


Plate 3. (Chapter on Properties of Interplanetary Dust by Jessberger et al., p. 279). TOF-SIMS secondary ion images of the original surface of U2071H9, an (Fe,Ni)S rich IDP. The halogens F, Cl, and Br are detectable on the very surface (from Rost et al., 1999).

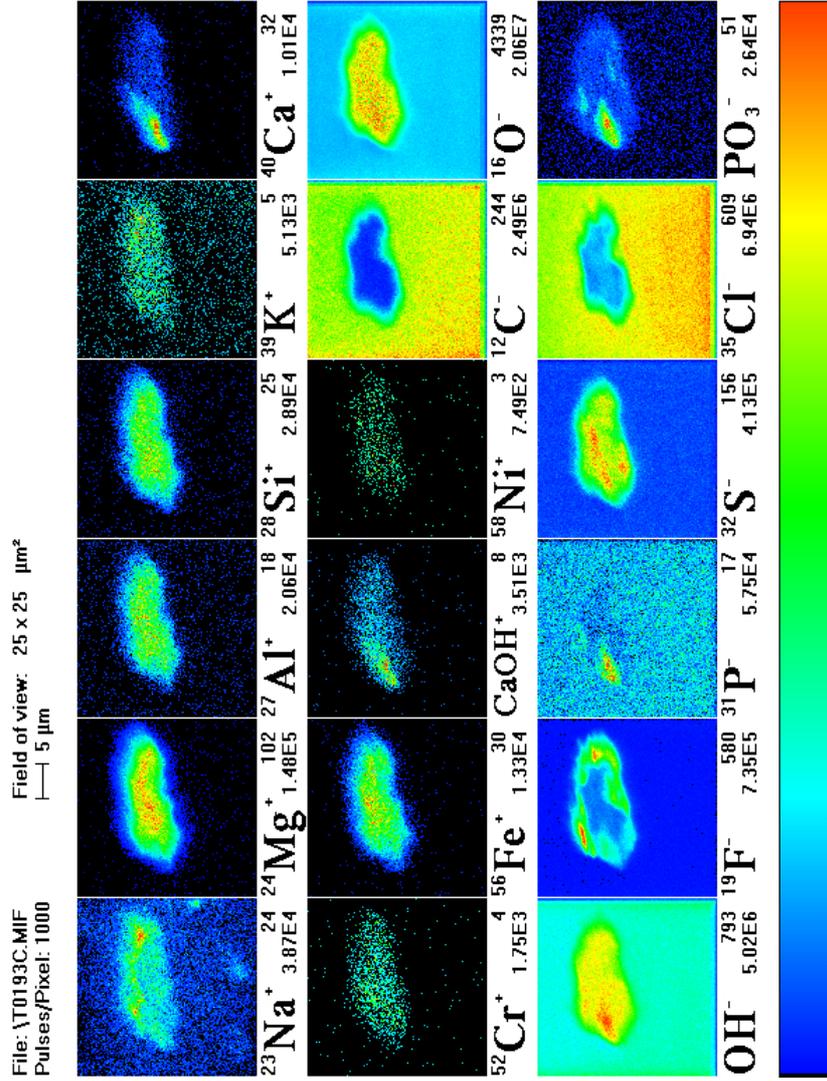


Plate 4. (Chapter on Properties of Interplanetary Dust by Jessberger et al., p. 280). TOF-SIMS secondary ion images (for explanation, see Plate 2) of the IDP L2006E10 reveal a Ca and P rich area, most probably an apatite grain.