The classification of micrometeorites

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Abstract–Due to their small size, the mineralogical and chemical properties of micrometeorites (MMs) are not representative of their parent bodies on the centimeter to meter scales that are used to define parent body groups through the petrological study of meteorites. Identifying which groups of MM are derived from the same type of parent body is problematic and requires particles to be rigorously grouped on the basis of mineralogical, textural, and chemical properties that reflect the fundamental genetic differences between meteorite parent bodies, albeit with minimal bias towards preconceived genetic models. Specifically, the interpretation of MMs requires a rigorous and meaningful classification scheme. At present the classification of MMs is, however, at best ambiguous. A unified petrological-chemical classification scheme is proposed in the current study and is based on observations of several thousand MMs collected from Antarctic ice.

INTRODUCTION

Micrometeorites (MMs) are extraterrestrial dust particles that have been captured by the Earth and are thought to mainly represent samples of asteroids and comets. In interplanetary space dust particles are inherently transported sunwards by P-R light drag (Dohnanyi 1967) and once captured by the Earth their deceleration occurs at high altitudes allowing fragile micrometeoroids to survive atmospheric entry (Love and Brownlee 1991). The parent bodies of MMs, therefore, are likely to differ from those of meteorites. Enormous numbers of particles can be recovered (up to tens of thousands during a single collecting expedition) (Maurette et al. 1991; Taylor et al. 1998), and provide us with a unique opportunity to study a diverse sample of solar system small bodies in the laboratory.

Studies of micrometeorites recovered from the Earth's surface and interplanetary dust particles (IDPs) collected in the stratosphere have shown that these include two distinct populations of materials. The larger micrometeorites (~30–1000 μm) recovered from the Earth's surface, which are mainly collected by filtering of melted Antarctic ice and snow (Duprat et al. 2003; Iwata and Imae 2001; Maurette et al. 1991; Taylor et al. 1998; Yada and Kojima 2000) mostly have strong mineralogical and chemical affinities with the carbonaceous chondrites, and thus appear to be derived principally from asteroids (Genge et al. 1997a; Kurat et al. 1994) although cometary materials are also present (Engrand and Maurette 1998a; Nakamura et al. 2005). Meteorite-like materials also occur among the smaller particles (<30 μm) collected directly from the Earth's stratosphere, however, around 40% of IDPs, known as anhydrous IDPs, are very different from meteorites and are highly volatile-rich. These have been suggested to be derived from comets (Bradley 1994; Rietmeijer 1998). Extraterrestrial dust particles collected in the stratosphere have come to be known as interplanetary dust particles, whereas those recovered from the Earth's surface are most commonly referred to as micrometeorites. Previously, only particles that survived atmospheric entry without melting were termed MMs (Whipple 1951); however, because of the gradational nature of melting, all extraterrestrial dust particles that survive to reach the Earth's surface, including melted particles, are here classified as MMs.

The small size of extraterrestrial dust particles restricts the information that individual particles can provide on the nature and origins of their parent bodies. Meteorites indicate that even the most primitive are relatively heterogeneous on

scales equivalent to the dimensions of MMs since they are composed of chondrules, refractory inclusions, and fine-grained matrix. Individual asteroidal MMs, therefore, provide us with samples of the components of their parent bodies, whether these are matrix, chondrules or refractory inclusions (Genge et al. 1997a; Kurat et al. 1994). In contrast, although anhydrous IDPs are mineralogically relatively homogeneous at scales larger than tens of microns (Rietmeijer 1998), the nature of cometary materials is uncertain.

The challenge with MMs, therefore, is that their mineralogical and chemical properties are unlikely to be representative of their parent bodies on the centimeter to meter scales that are used to define parent body groups through the petrological study of meteorites (Brearley and Jones 1998). Identifying which groups of MM are derived from the same type of parent body is problematic and requires particles to be rigorously grouped on the basis of mineralogical, textural, and chemical properties that reflect the fundamental genetic differences between meteorite parent bodies, albeit with minimal bias towards preconceived genetic models. Specifically the interpretation of MMs requires a rigorous and meaningful classification scheme.

Here we present a unified classification system for MMs intended to reflect the genetic relationships between different groups of MMs that may derive from related parent bodies. The classification system is based on the combined observations of several thousand particles recovered from Antarctic ice from both the vicinity of Cap Prudhomme (Maurette et al. 1991) and the South Pole Water Well (Taylor et al. 1998) by the authors. Descriptions of the techniques used to select, prepare, and analyze these materials are given in Engrand et al. (1998a), Genge et al. (1997a), and Taylor et al. (2000). Groups of MM proposed herein are all based on backscattered imaging of polished grain mounts of particles and, in some cases, quantitative microanalysis of mineral phases, however, throughout this paper we also provide descriptions of features by which MM types can be approximately identified under a binocular microscope during the selection procedure.

NON-ISOTOPIC CRITERIA FOR EXTRATERRESTRIAL ORIGIN

Definitive evidence for the extraterrestrial origin of MMs has been made on the basis of cosmogenic nuclei and noble gas measurements (Olinger 1990; Raisbeck and Yiou 1987); however, the identification of dust particles recovered from the Earth's surface as micrometeorites can be made on one or more of a number of criteria. Features that strongly suggest an extraterrestrial origin are any of: (1) the presence of a partial or complete shell of magnetite around MMs, which is thought to arise from entry heating (Toppani et al. 2001; Toppani and Libourel 2003), (2) the presence of Ni-bearing iron metal, and (3) a chondritic

bulk composition for major and minor elements, at least for particles with a small grain-size relative to particle-size. Features that are less determinative are: (1) high CaO, Cr_2O_3 olivines and very FeO-poor olivines that are exceedingly rare in terrestrial rocks (Brearley et al. 1998), (2) evidence for surface heating consistent with atmospheric entry, and (3) spherical particle morphologies.

Spherical particle morphologies are particularly ambiguous since anthropogenic particulates, impact spherules, meteorite ablation spheres and even some small-scale volcanic dust may have similar morphologies. Iron-rich industrial/diesel spherules are, for example, very common in the terrestrial environment and many consist primarily of magnetite and are thus mineralogically indistinguishable from some melted micrometeorites. Isotopic data of iron-rich spherules recovered from the deep sea have confirmed their extraterrestrial origins (Clayton et al. 1986; Engrand et al. 1998, 2005; Herzog et al. 1999; Raisbeck and Yiou 1989). These make up a large fraction of the deep sea spherules (25 to 50%), but are present in abundances of a few percent in polar regions (Genge et al. 1997a; Taylor et al. 1998). In early collections in continental areas outside the polar regions, iron-dominated spherules were found to be abundant, and studies have shown these are of anthropogenic origin (Fisher et al. 1976).

METEORITE ABLATION SPHERES AND MICROTEKTITES VERSUS MICROMETEORITES

Meteorite ablation spheres, produced during the atmospheric entry of larger meteoroids, and microtektites, condensed from vapor plumes or separated from impact melt produced during the collision of asteroids and comets with the Earth, could also be considered extraterrestrial in origin. In the case of microtektites a small extraterrestrial component is intimately mixed with terrestrial materials from target rocks (Glass 1990). Micrometeorites are extraterrestrial particles that were present as dust-sized particles in interplanetary space and were not part of larger meteoroids. Spallogenic and cosmogenic isotope abundances consistent with exposure to solar energetic particles and galactic cosmic rays as small particles allow MMs to be unequivocally distinguished from ablation spherules and microtektites (Harvey et al. 1998; Raisbeck and Yiou 1987, 1989). However, there are also several mineralogical and chemical features that can be used to distinguish these groups of materials in routine particle surveys.

For unmelted dust-sized debris the presence of a magnetite-rim unequivocally distinguishes MMs since this is developed during hypervelocity deceleration at high altitudes (Toppani et al. 2001; Topponi and Libourel 2003). Discrete magnetite rims are not present either on meteorite fusion crusts (Genge et al. 1999), ablation spheres (Harvey et al.

1998), or on microtektites (Glass 1990). However, MMs that have experienced a high degree of melting during atmospheric entry heating (e.g., cosmic spherules) or very low degrees of heating likewise generally lack magnetite rims.

Studies of meteorite fusion crusts (Genge et al. 1999) and spherules from the one known meteorite ablation spherule layer (Harvey et al. 1998) suggest that ablation spherules will have bulk compositions very close to that of their parent meteoroid and are enriched in volatile and moderately volatile elements such as Na and S relative to most melted MMs, which have lost volatiles by evaporation. Microtektites usually consist of a mixture of target and projectile materials and thus have major element compositions that diverge significantly from chondritic (Glass 1990), allowing them to be distinguished from MMs. Metallic impact spherules, associated with iron projectiles and commonly associated with smaller craters such as the Barringer crater are, however, broadly similar in composition to iron-rich melted MMs, except in their Cr contents.

Probably the most significant feature of meteorite ablation spheres and impact spherules, however, is their localized temporal, spatial, and chemical distribution. The Bit-58 meteorite ablation layer in the Antarctic, for example, contains spherules with a very limited range of chemical and isotopic characteristics related to a H-chondrite parent body (Harvey et al. 1998). Close to Barringer Crater likewise iron-spherules are found in much higher abundances than iron-rich melted MMs (Nininger 1956). Silicate dominated impact spherules, associated with larger impacts, are stratigraphically restricted. Impact and ablation spherules can, therefore, be distinguished on the basis of their localized distribution.

CLASSIFICATION OF MICROMETEORITES

Two main groups of micrometeorite can be identified on the basis of their preatmospheric textures: (1) fine-grained MMs (FgMMs), which are dominated by a fine-grained porous groundmass of micron-sized mineral grains, and (2) coarse-grained MMs (CgMMs), which are dominated by anhydrous silicates with grain-sizes larger than several microns, often with glassy mesostasis. Heating of MMs during atmospheric entry, however, complicates the classification of particles on the basis of their preatmospheric nature because it results in significant changes in the primary mineralogy, texture and even the compositions of particles. Micrometeorites are, therefore, also divided into several groups (Table 1) depending on the extent of thermal reprocessing during atmospheric entry. The proportion of melted, partially melted, and unmelted micrometeorites is not well known but varies with particle size (Maurette et al. 1991; Taylor et al. 2000). For sizes >100 μm melted MMs make up 70–90% of particles (Maurette et al. 1991; Taylor et al. 2000), for those 50–100 mm in size melted and partially melted

MMs make up ~50% of particles (Genge et al. 1997a), although a partially melted MMs to unmelted ratio of 1 in this size range is found using a different classification (Engrand and Maurette 1998), and for those 25–50 mm in size melted and partially melted MMs comprise only 22% of the particles (Gounelle et al. 2005b). It is stressed that changes due to atmospheric heating are largely gradational and, therefore, assigning micrometeorites into a discrete group is not always possible. This has lead to differences in how MMs have been classified.

Melted Micrometeorites: Cosmic Spherules

Melted micrometeorites are classified as spherical to subspherical particles formed as molten droplets during atmospheric entry. These particles that have experienced large degrees of fusion of primary phases (defined in the current manuscript as those existing prior to atmospheric entry) during atmospheric entry and thus behave as low viscosity melts (Fig. 1). Igneous particles that experienced melting prior to capture by Earth are thus not included in this group and are classified as unmelted MMs. It is, however, not always straightforward to distinguish particles that have melted in the atmosphere from those that were primarily melted.

Evidence that significant fusion occurred during atmospheric entry is an important criteria for identifying a melted MM as compared to an igneous unmelted particle. The fundamental characteristic of atmospheric entry heating is that it is a surface process. Surface correlated heating or particle shapes suggesting melting with no subsequent thermal overprint thus strongly imply that heating occurred during atmospheric entry. Fusion is identified in melted MMs on the basis of textures and mineralogy. Particles dominated by a glassy mesostasis, with or without microphenocrysts, are demonstrably formed by crystallization of a melt. Vesicles are also often present within the mesostases of melted MMs, however, they are not necessarily characteristic of these particles since subspherical voids also occur in unmelted varieties.

Cosmic spherules show considerable diversity in textures, compositions and mineralogy and can be subdivided into several chemical and textural groups (Fig. 1). The basic chemical subtypes of CSs, which are also reflected in their principle mineralogy, are the iron-rich spherules (I-type), a glass with magnetite (G-type) group (Blanchard et al. 1980) and silicate-type (S-type) CSs.

I-type CSs (Figa. 1a and 1b) are dominated by FeO with minor amounts of other oxides (principally MgO and SiO^2), mineralogically they are dominated by the iron oxides wüstite and magnetite, however, Ni-rich iron metal, sometimes with nugget of the platinum group elements, can occur as spheres within spherules. I-type spherules often contain a single large spherical void that may form by the

Table 1. An outline of the classification of micrometeorites.

Groups	Class	Type	Subtype	Description
Melted MMs	Cosmic spherules (CSs)	S	CAT	Spherules with Mg/Si > 1.7 that are enriched in Ca, Ti, and Al. They have barred olivine textures.
		S	Glass	Spherules consisting almost entirely of glass.
		S	Cryptocrystalline	Spherules dominated by submicron crystallites and magnetite. Some include multiple domains.
		S	Barred olivine (BO)	Spherules dominated by parallel growth olivine within glass.
		S	Porphyritic olivine (Po)	Spherules dominated by equant and skeletal olivine within glass. Relict-bearing varieties contain unmelted minerals.
		S	Coarse-grained	These spherules contain >50% volume relict minerals.
		G		Spherules are dominated by magnetite dendrites within silicate glass.
		I		Spherules dominated by magnetite, wüstite.
Partially melted MMs	Scoriaceous MMs (ScMMs)	–	–	Vesicular particles dominated by a mesostasis of fayalitic olivine microphenocrysts within glass. ScMMs often contain relict minerals and relict matrix areas.
Unmelted MMs	Fine-grained MMs (FgMMs)	C1		Compact, chemically homogeneous FgMMs. Often contain framboidal magnetite.
		C2		Compact, chemically heterogeneous fine-grained MMs. Often contain isolated silicates and/or tochilinite.
		C3		Highly porous FgMMs. Often contain isolated silicates and framboidal magnetite.
	Coarse-grained MMs (CgMMs)	Chondritic CgMMs	Porphyritic olivine and/or pyroxene	Igneous MMs dominated by pyroxene and/or pyroxene phenocrysts within glass.
			Granular olivine and/or pyroxene	Igneous MMs dominated by pyroxene and/or olivine without significant glass.
			Barred olivine	Igneous MMs dominated by parallel growth olivine within glass.
			Radiate pyroxene	Igneous MMs dominated by radiating pyroxene dendrites within glass.
			Type I/type II	Type I CgMMs are reduced particles containing Fs and/or Fa < 10 mol%. Type II CgMMs are oxidized particles with Fs and/or Fa > 10 mol%.
		Achondritic CgMMs	–	Differentiated igneous CgMMs.
	Refractory MMs	Porous	–	Porous particles dominated by refractory minerals.
		Compact	–	Compact particles dominated by refractory minerals.
		Hydrated	–	Particles dominated by refractory minerals surrounded by Fe-rich phyllosilicates or their dehydroxylates.
	Ultracarbonaceous MMs		–	Particles dominated by carbonaceous materials with embedded

loss of a molten metal bead from the particle. They also often exhibit an irregular void space in their centre which could result from the rapid cooling of the melt from the surface inwards (Feng et al. 2005). I-type spherules are abundant in deep sea collections, as they are very resistant to weathering, but they constitute only 2% of 1600 cosmic spherules collected at the bottom of the South Pole Water Well (SPWW) (Taylor et al. 2000).

G-type CSs have major element compositions that are intermediate between the broadly chondritic S-type spherules and the I-type CSs. G-type spherules are typically dominated by magnetite dendrites within a mesostasis of silicate glass, some have subspherical voids similar to those found within

I-type spherules which may likewise have formed by the loss of metal beads (Figs. 1c and 1d). G-type spherules, however, can have a wide range of textures and include particles that differ in texture and mineralogy from typical spherules. G-type spherules are only present at the level of 1% among 1600 CSs from the SPWW (Taylor et al. 2000).

The silicate S-type are by far the most common, making up 97% of 1600 cosmic spherules analyzed from the South Pole Water Well (Taylor et al. 2000). Most have broadly chondritic compositions (Brownlee et al. 1997); notable exceptions are the CAT spherules that have Mg/Si ratios >1.7 and are highly enriched in Ca, Al, and Ti (Taylor et al. 2000). S-type CSs are dominated by olivine microphenocrysts,
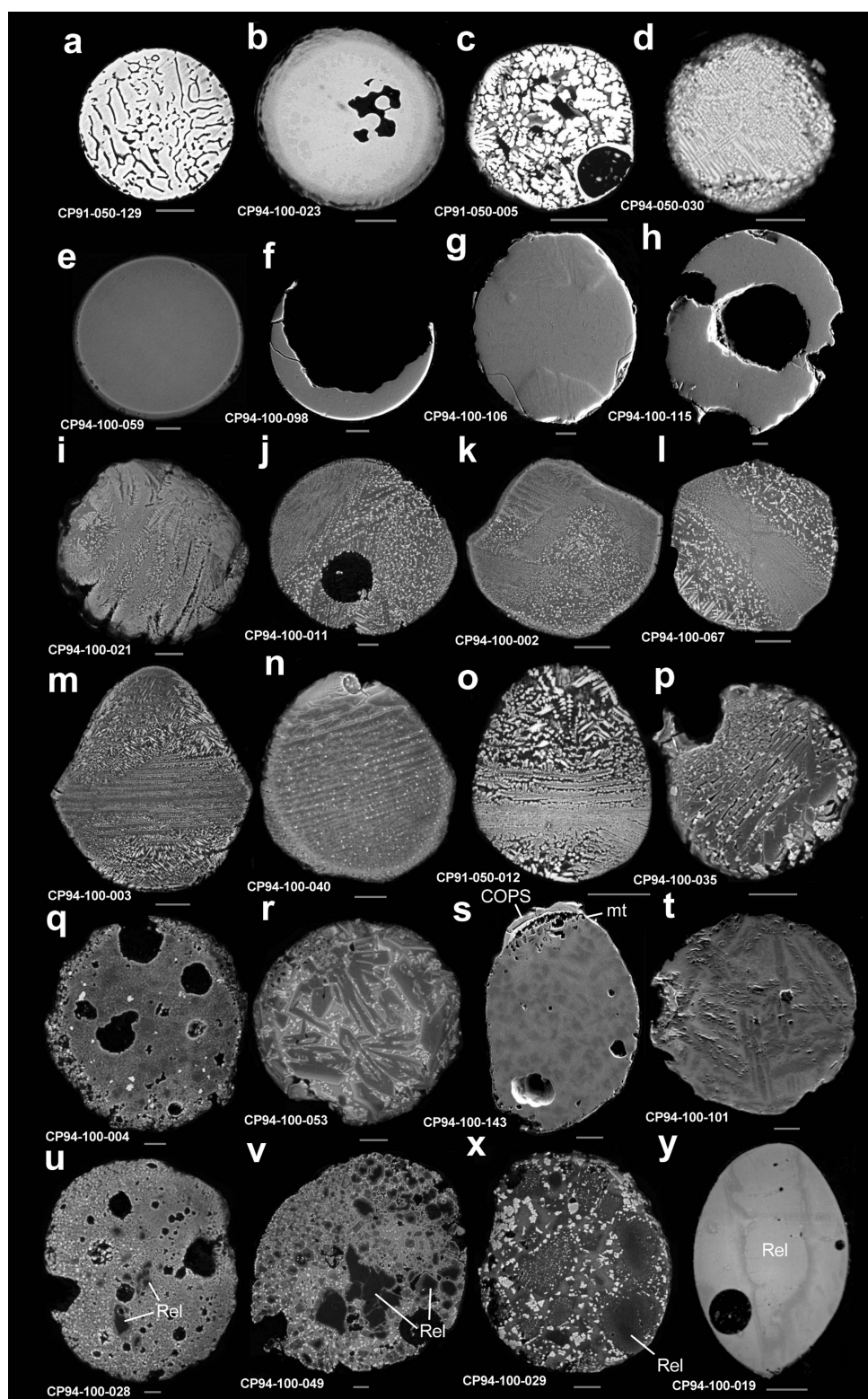
Fig. 1. Backscattered electron images of cosmic spherules. a), b) I-type spherules dominated by magnetite and wüstite. c), d) G-type spherules containing magnetite dendrites in a silica-rich glass. e–h) Glass S-type spherules. i–k) Cryptocrystalline S-type spherules. l–p) Barred olivine S-type spherules, where l is probably transitional to cryptocrystalline spherules. (q–t) Porphyritic olivine S-type spherules. Spherule s has a magnetite rim (mt) associated with COPS. u), v) Relict-bearing PO spherules containing relict anhydrous silicates (Rel) that have survived atmospheric melting. x), y) Coarse-grained spherules, which are dominated by relict grains (Rel). In particle x some relict contain iron-oxide inclusions. The scale bar is 20 μm.

silicate glass and often contain magnetite and/or chromite. They may also contain relict grains, principally Mg-rich pyroxene and olivine, that survived melting in the atmosphere and sometimes have FeNi metal droplets.

S-type spherules can be subdivided into several subclasses depending on their quench textures, which are thought to reflect their peak atmospheric temperatures (Taylor et al. 1991). CAT spherules, have barred olivine textures, lack Fe, have high Mg/Si ratios, and high Ca, Al, and Ti values. These appear white under the microscope and weather quickly in the acidic Antarctic melt water (Taylor et al. 2000). Isotopic analyses (Mg, Si, and O) confirm that they have been partial evaporated during atmospheric entry heating (Alexander 2002) and compared to all other S-type spherules were subjected to the highest temperatures.

Glass spherules (Figs. 1e–h) lack olivine microphenocrysts and are thought to have formed at the next highest peak temperatures. Under a binocular microscope these tend to be transparent brown, yellow, or green in color and go extinct under crossed polarization. Glass spherules sometimes contain large vesicles. They are usually spherical, can be highly vesiculated, and some contain FeNi metal beads. They should not be confused with the G-type spherules that were described above.

Cryptocrystalline (CC) spherules (Figs. 1 i–l) contain submicron crystallites and can have significant submicron magnetite. There are two distinct textures. Olivine microcrystals can grow from the surface inward producing a knobbly surface. Under a binocular microscope the knobbly protrusions on the surface of CC spherules are characteristic and were called "turtleback" (Brownlee and Bates 1983). Other cryptocrystalline spherules have multiple regions that crystallized simultaneously producing extremely fine-grained regions surrounded by a more iron rich phase. CC spherules are thought to have experienced slightly lower peak temperatures than the glass spheres, because some of crystallization nuclei survive (Brownlee et al. 1991).

Barred olivine (BO) spherules (Figs. 1m–p) are dominated by parallel growth olivine, which is seen as parallel bars in polished sections, within a glassy mesostasis that often contains magnetite. Most BO spherules have ovoid (egg-like) shapes that are easily identified under a binocular microscope, although some are elliptical. In ovoid BO spherules, olivine bars are usually orientated approximately perpendicular to the long-axes of the particles. Some BO spherules exhibit FeNi metal beads that have sometimes oxidized to form iron-oxide, often with a cubic morphology, located at one end of the particle. The Fe-Ni-oxides frequently contain S, C, and P, and are probably dominated by ferrihydrite (Engrand et al. 1999) and might represent weathering products of metal (Genge et al. 1998a). These iron-rich bodies, therefore, probably formed as immiscible Fe-Ni-metal liquids that were separating from the silicate

melt of the spherule (Genge et al. 1998a). Barred olivine spherules are thought to form at lower peak temperatures than CC spherules (Taylor et al. 2000). Barred olivine spherules are usually opaque under a binocular microscope, except for those containing a metal/ferrihydrite bead which are often brown/green transparent.

Porphyritic olivine (PO) spherules (Figs. 1q–t) are dominated by olivine microphenocrysts with equant, euhedral or skeletal morphologies within a glassy mesostasis, usually with accessory magnetite and/or chromite. Relict olivine (and rare pyroxene), often with overgrowths of more Fe-rich olivine grown from the melt, are common in PO spherules and together with the olivine microphenocrysts, which indicate abundant crystallization nuclei, suggest that PO spherules experienced the lowest peak temperatures of any cosmic spherules, especially those that contain relict minerals.

Porphyritic spherules show a wide range of crystallinities and crystal sizes. Those with the smallest microphenocrysts tend to be the most vesicular and are likely to be gradational to partially melted MMs. Some PO spherules also contain areas dominated by Fe-Ni-ferrihydrite and/or Ni-bearing sulfides. These are thought to form as immiscible metallic liquids during heating and are often found at the margins of spherules, suggesting they were in the processes of separating during cooling (Genge et al. 1998a).

Relict pyroxene and olivine within cosmic spherules that have survived atmospheric melting are usually forsterite and enstatite, however, occasionally Fe-rich varieties are observed and identified by more Mg-rich overgrowths. The abundance of relict grains varies greatly from those containing small volumes of isolated grains, often present as cores to olivine microphenocrysts grown from the melt, to spherules in which relicts are volumetrically dominant. The latter particles probably represent melted coarse-grained or composite particles (see below for definition of these materials) and are, thus, distinct from other CSs that form by melting of fine-grained particles.

The presence of relict minerals within CSs should be denoted by a prefix of "relict-bearing" (Taylor et al. [1998] and Figs. 1u and 1v). Those spherules that appear to have formed by melting of a coarse-grained precursor should be described as CG spherules. These spherules include those containing large volumes (>25%) of relict grains or having non-chondritic compositions suggesting formation by melting of a coarse-grained precursor (Figs. 1x and 1y).

Partially Melted MMs

Scoriaceous Fine-Grained MMs

Scoriaceous micrometeorites (ScMMs) are irregular, but smooth, highly vesicular particles (Fig. 2). Particle shape and vesicularity alone are not, however, particularly

discriminative properties for these particles since they represent a gradational series from the more intensely heating CSs to the less heated ScMMs. The presence of envelopes of magnetite (magnetite rims) surrounding ScMMs and their absence on CSs, however, seems to be much more characteristic even if the reason for this difference is, as yet, largely uncertain (Toppani et al. 2001).

Scoriaceous micrometeorites are dominated by a mesostasis of microporphyritic olivine, usually with crystal sizes <1 µm, within an interstitial silicate glass phase (Fig. 2). Although the textures of the mesostasis implies crystallization of a melt, probably at temperatures just above the melting point, ScMMs commonly contain areas of relict fine-grained matrix similar to that of many unmelted MMs. Many ScMMs are, therefore, partially melted and are a gradational group to unmelted particles. Under a binocular microscope ScMMs often have lobate, smooth exteriors with vesicles impinging on their surfaces.

The most common relict grains in ScMMs and are Mg-rich pyroxene and olivine. Enstatite grains are frequently highly fractured and are surrounded by more Fe-rich pyroxene that appears to have formed by reaction with the surrounding melt. Relict olivines are usually forsterites. Vesicle abundances in ScMMs are frequently high and sometimes exceed 50% by volume. Occasionally ScMMs contain Fe-Ni metal/oxides or Ni-bearing sulfides and like those of CSs these may have formed as immiscible metallic/sulfide liquids.

One ScMM has been reported (Genge and Grady 1998a) with large volumes of Fe-Ni-oxides and a mesostasis dominated by enstatite microphenocrysts rather than olivine. No additional similar ScMMs have yet been recognized, however, it remains a possibility that there are a subgroup of Mg-rich ScMMs formed by separation of significant Fe-Ni-metal liquids under highly reducing conditions. Such particles may have formed from C-rich precursors.

Distinguishing partially melted mesostasis and relict matrix is difficult since in reality both have probably experienced a degree of fusion. Partially melted mesostasis, however, is probably best defined as areas that have melted to a sufficient degree during entry heating to behave as a viscous fluid, whereas relict matrix retains the rheological properties of a rigid solid. Partially melted mesostasis, therefore, lacks irregular pore space, except that which can be attributed to etching of interstitial glass, and irregular dehydration cracks. Both partially melted mesostasis and relict matrix, however, contain subspherical to rounded vesicles, although these tend to have more irregular shapes within relict matrix. The presence of Fe-rich olivine microphenocrysts within the partially melted mesostasis of ScMMs also gives it a higher backscattered electron potential than relict matrix (Figs. 2 and 3).

Studies of the thermal evolution of micrometeoroids during entry heating (Flynn 1995; Genge 2006) and heating experiments (Toppani et al. 2001) indicate that the survival of relict matrix within ScMMs occurs due to thermal discontinuities supported by decomposition of volatile-bearing phases such as phyllosilicates and carbonaceous matter. Consequently the fusion of FgMMs occurs through surface melting and is a gradational process in which the unmelted core is progressively consumed.

In the current classification system it is proposed that ScMMs with relict fine-grained cores that comprise less than 50% by volume of the particle are termed Sc/FgMMs; where Fg denotes fine-grained matrix. Particles in which the relict fine-grained core exceeds 50% by volume of the particle are denoted as Fg/ScMM particles.

The matrix mineralogies and textures of relict matrix are described in detail below since thermal alteration of fine-grained matrix is a gradational and complex process that is likely to be dependent on the duration of heating as well as peak temperature. Consequently thermally altered, relict matrix is also often observed in particles lacking scoriaceous mesostasis.

Scoriaceous Coarse-Grained MMs

Scoriaceous coarse-grained MMs are formed by partial melting of CgMMs, which are described below amongst unmelted particles. Incorporating the degree of thermal alteration within CgMMs in a classification scheme is complicated due to the gradational nature of thermal alteration affects within these materials. The lowest temperature changes in mineralogy occur in the glassy mesostases of particles with the remobilization of the glass, the formation of vesicles and growth of fayalitic olivine overgrowths on olivine phenocrysts (Genge et al. 1996). Partial melting occurs at temperatures above the solidus and can produce CSs. Igneous rims found on many CgMMs may be melted fine-grained matrix that was attached to the exterior of the particle (Genge 2006) and provide a direct relationship to the thermal alteration of fine-grained particles. As defined for scoriaceous/fine-grained particles, scoriaceous MMs which have retained an large (>50%) fraction of unmelted coarse-grained minerals will be noted as Cg/ScMMs (Figs. 9a and 9b). If the particle shows large amounts of scoriaceous matrix associated with coarse-grained minerals, the MM will be described as Sc/CgMM (Fig. 6).

Unmelted Micrometeorites

Unmelted fine-grained MMs are those dominated by a fine-grained porous groundmass of micron-sized mineral grains and are similar to the fine-grained matrices of chondritic meteorites (Figs. 3 and 4). Like these materials they have broadly chondritic compositions, mostly in the range of CI, CM, and CR chondrite matrices, for most major and minor elements (Genge et al. 1997a; Kurat et al. 1994). The depletions in Ca, Ni, and S present within Cap
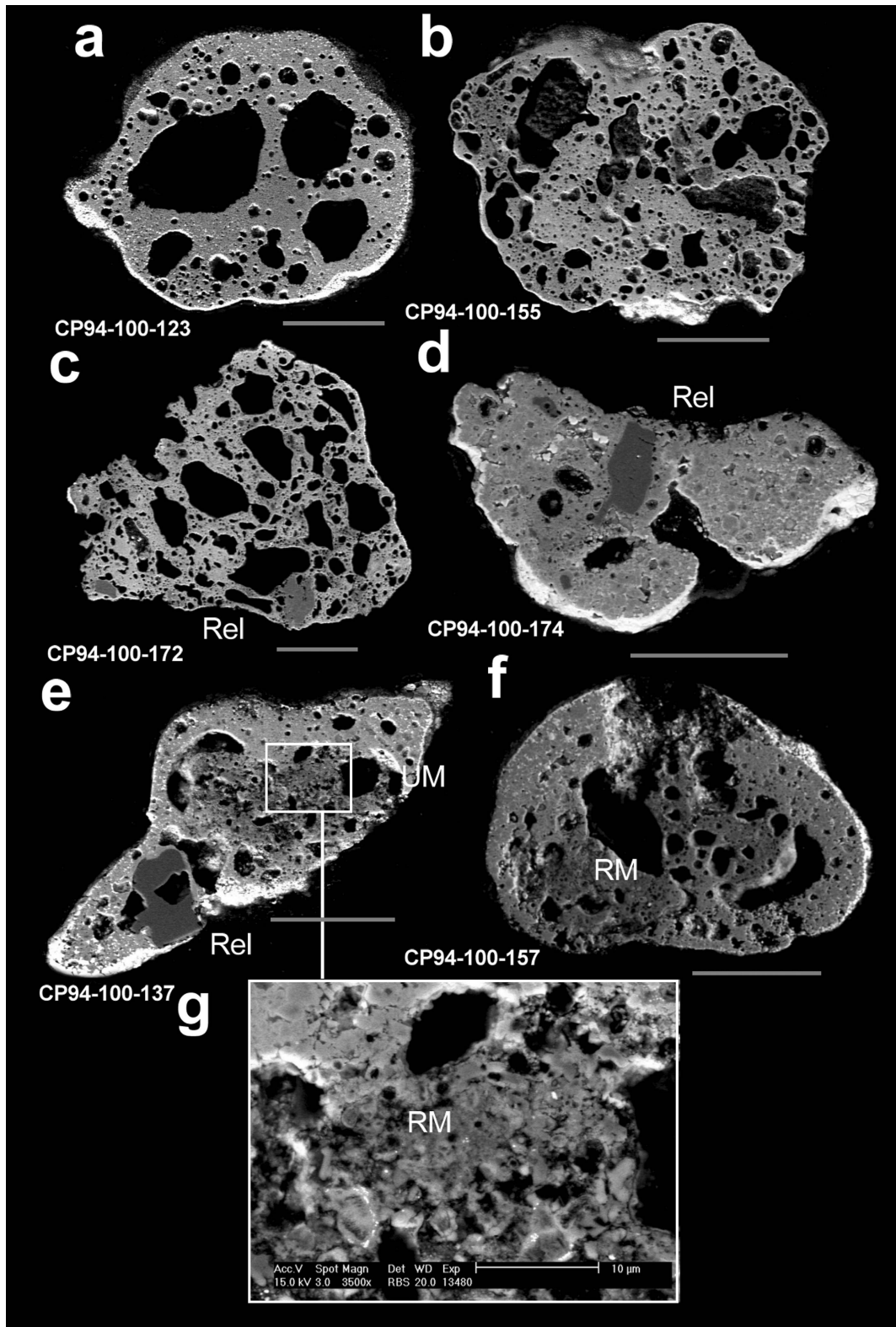
Fig. 2. Backscattered electron images of scoriaceous micrometeorites. Particles c and d are relict-bearing (Rel) since they contain anhydrous silicates (forsterite and enstatite) that have survived atmospheric heating. Particles e and d both contain cores of relict fine-grained matrix (RM). g) A high-resolution backscattered electron image of particle e showing the melted mesostasis of the ScMM (top of image; dominated by fayalitic olivine microphenocrysts) and unmelted matrix (center of image; dominated by lower backscattered potential porous materials). The scale bar is 50 μm.
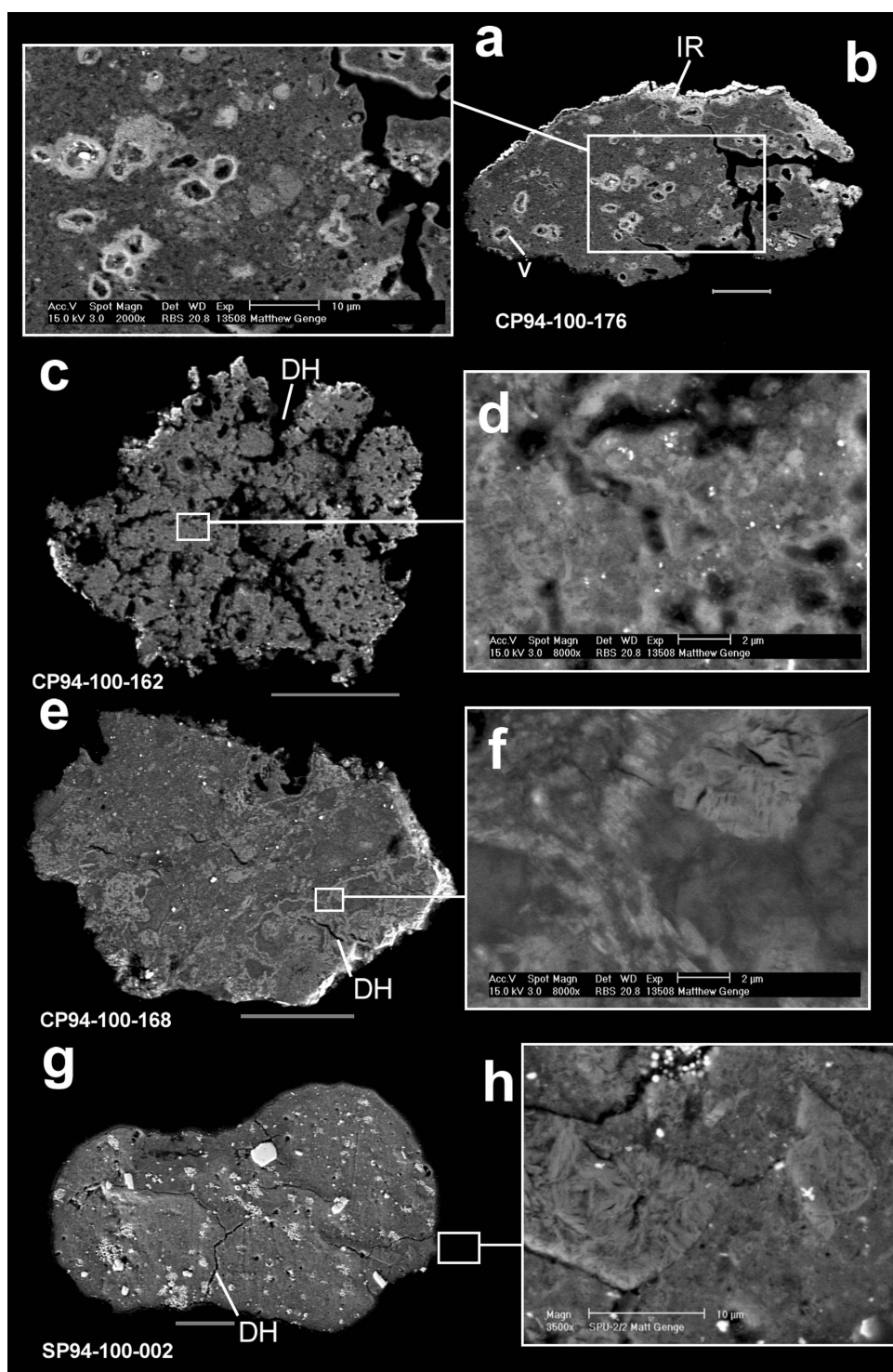
Fig. 3. Backscattered electron images of unmelted micrometeorites. a), b) An unmelted MM with an igneous rim (IR) consisting of microphenocrysts of fayalitic olivine within a glassy mesostasis surrounding unmelted, but altered, matrix with vesicles. c), d) An altered MM dominated by fine-grained matrix lacking textural evidence for acicular/sheet-like phases but with irregular dehydration cracks (DH). This particle also has a thin igneous rim and is transitional to ScMM. e), f) An unmelted MM consisting of fine-grained matrix with textural evidence for acicular/sheet-like phases. Transmission electron microscopy of this particle (Genge et al. 2001) indicates that phyllosilicates have decomposed to amorphous dehydroxylates. g), h) An unmelted MM consisting of fine-grained matrix with textural evidence for acicular/ sheet-like phases. Transmission electron microscopy of this particle confirms the presence of serpentine and saponite. The presence of unaltered phyllosilicates cannot be determined from textures observed in backscattered electron images, however, the absence of a magnetite rim may be indicative of its low degree of heating. Scale bar is 50 μm.
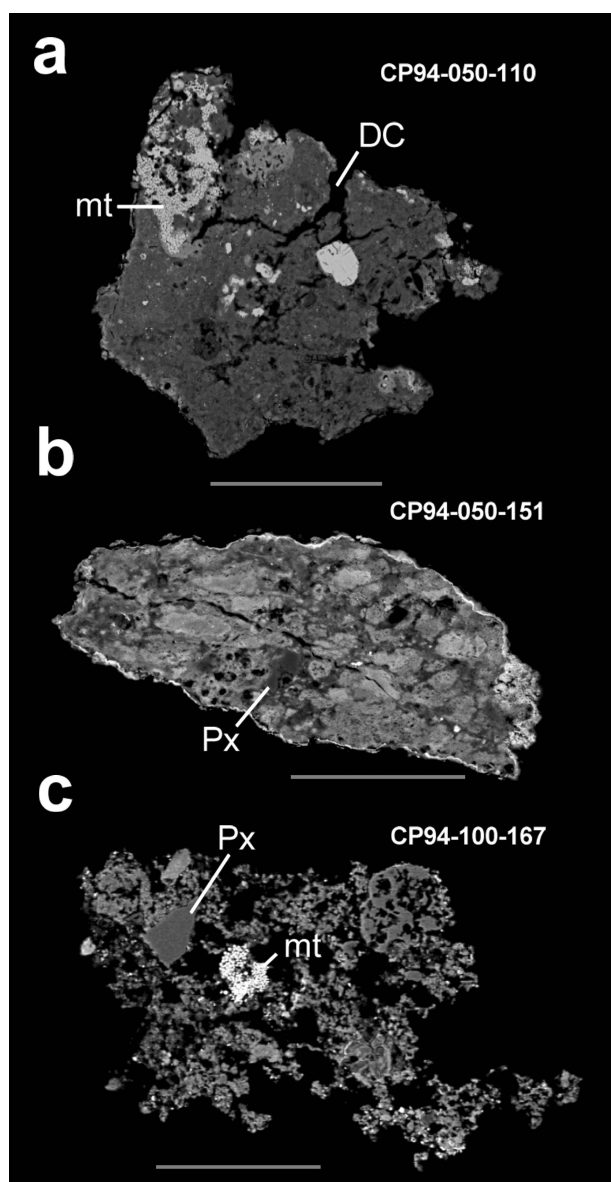
Fig. 4. Backscattered electron images of textural types of MM suggested by Genge et al. 2002a. a) A compact C1 MM containing framboidal magnetite (mt) and showing dehydration cracks (DC); Fig. 3g also shows a C1 particle. b) A compact C2 MM with significant spatial variation in the Fe/Mg ratio of the matrix. This particle contains a small isolated silicate mineral (IS), in this case pyroxene. Figure 3e also shows a C2 particle. c) A porous C3 MM dominated by anhydrous silicates with framboidal magnetite. The matrix of the composite particle in Fig. 7 is also C3 material. Scale bar is 50 µm.

Prudhomme and SPWW micrometeorites may be in part due to the leaching of soluble phases like sulfides and carbonates while in the ice/water, as they are not found in Concordia MMs (Duprat et al. 2005). Primary differences from chondrites, for example, in the high pyroxene to olivine ratios of MMs are also evident (Maurette et al. 1991).

Unmelted coarse-grained micrometeorites are defined as dominated by anhydrous silicates, primarily pyroxene with subordinate pyroxene, with grain sizes larger than ~1 micron. Often CgMMs have igneous textures with olivine and/or pyroxene contained within a glassy mesostasis, however, some CgMMs are merely fragments of single crystals of olivine or pyroxene, or are granular aggregates of anhydrous silicates (Fig. 5).

Composite unmelted micrometeorites have also been observed that contain both portions characteristic of both CgMM and FgMM (Genge 2006). Such particles usually consist of a coarse-grained, igneous-textured object dominated by anhydrous silicates and glassy mesostasis, surrounded by a partial rim of fine-grained matrix (Fig. 7). To denote the composite nature of these particles the notation Fg/CgMM and Cg/FgMM is used to describe particles in which the fine-grained and coarse-grained portions dominate respectively.

Unmelted MMs: Fine-Grained MMs

Thermal Alteration of Hydrous Matrix in Fine-Grained MMs

Genge et al. (1997a) suggested that FgMMs could be split into altered and unaltered subtypes due to thermal alteration during atmospheric entry in which unaltered particles are those containing acicular to sheet-like, submicron mineral grains in the matrix. These were suggested to be phyllosilicates or their thermal decomposition products. Particles whose matrices are dominated by submicron anhedral, equant grains, often within a interstitial homogeneous material were defined as altered and suggested to consist of matrix in which the amorphous dehydroxylates of phyllosilicates that have decomposed into olivine, pyroxene, and glass (Greshake et al. 1998). Such particles frequently contain subspherical vesicles that are probably formed by decomposition of volatile-bearing phases and often have rims of redeposited amorphous material (Genge et al. 2001).

Transmission electron microscope (TEM) and X-ray diffraction studies of the matrices of FgMM have confirmed that phyllosilicates with measurable basal spacings are rare and are dominated by smectite (Genge et al. 2001; Gounelle et al. 2002; Nakamura et al. 2001; Noguchi et al. 2000, 2002) although serpentine has also been identified. Most FgMMs whose matrices contain submicron acicular to sheet-like phases are dominated by amorphous, often iron-rich silicates. These appear to be pseudomorphs after original hydrous minerals and are thought to be dehydroxylates after phyllosilicates (Genge et al. 2001; Gounelle et al. 2002; Noguchi et al. 2000).

Transmission electron microscope studies of more intensely heated FgMMs and the relict fine-grained matrices of ScMM indicate they are dominated by submicron, anhedral olivine and pyroxene with interstitial glass and thus probably are the thermal decomposition products of dehydroxylates. It is likely, however, that the transition from phyllosilicates-

bearing MMs to those in which amorphous dehydroxylates of phyllosilicates have recrystallized to olivine and pyroxene is gradational. Genge et al. (2001) suggested FgMMs lacking magnetite rims are probably the least heated and most likely to contain undecomposed phyllosilicates.

Many FgMMs preserve the original texture of the fine-grained matrix to a large extent since the amorphous dehydroxylates pseudomorph the pre-existing phyllosilicates. Those FgMMs that preserve phyllosilicates with measurable basal-spacings are the least heated and should be prefixed with phyllosilicate-bearing.

Primary Variations

It has been suggested that FgMMs may also be subdivided on the basis of their textures and chemical homogeneity into one of three groups (Genge and Grady 2002): CU1, CU2, and CU3 FgMMs. Chemical homogeneity in Fe/Si ratio, which is reflected in significant variation in backscattered electron potential, in particular can also be distinguished in thermally altered FgMMs (Genge et al. 2000).

CU1 particles are compact FgMMs with low apparent porosities that are chemically homogeneous over scales of 10 microns (Fig. 4a). CU1 particles rarely contain large (>4 μm), "isolated" anhydrous silicates. These particles are suggested to have affinities to carbonaceous chondrite type 1 meteorite matrices (i.e., those which have experienced intense aqueous alteration such as the CI1 chondrites and the Tagish Lake chondrite (Brearley and Jones 1998; Zolensky et al. 2002).

CU2 particles are compact FgMMs with low porosities that are chemically heterogeneous, in particular in their Fe/Mg, Fe/Si ratios, over distances of 10 microns (Fig. 4b). Isolated, anhydrous pyroxenes and olivines more than 4 μm across are more abundant in CU2 particles than in CU1. These particles may have affinities to carbonaceous chondrite type 2 meteorite matrices that, although aqueously altered, have not been chemically homogenized on small scales.

CU3 particles are porous FgMMs, sometimes with a porosity up to 50% by volume, that are dominated by subhedral to euhedral magnesian olivine and pyroxene grains, which are up to several microns in size, with minor amounts of interstitial acicular to sheet-like phases (Fig. 4c). A TEM investigation of one CU3 particle suggests the sheet-like phase is a dehydroxylate after phyllosilicate (Genge and Grady 2002). Based on the higher visible irregular porosity and the paucity of phases after hydrous silicates CU3 particles might have affinities with carbonaceous chondrite type 3 meteorite matrices that have experienced lower degrees of aqueous alteration than type 1 or type 2.

Under a binocular microscope FgMMs have irregular shapes and vary from smooth compact particles (generally CU1 and CU2) to highly re-entrant fluffy particles (CU3). In altered FgMMs subspherical voids can often be observed on their surfaces and particles with igneous rims can resemble ScMMs.
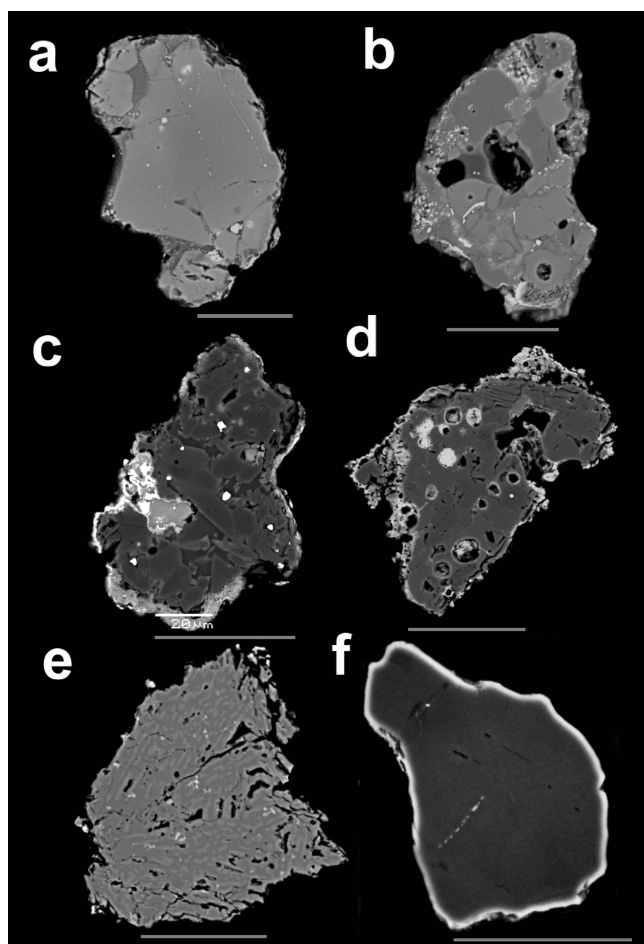

Fig. 5. Backscattered electron images of chondritic CgMMs. a) Porphyritic pyroxene type I CgMM with interstitial glass and FeNi droplets, some of which have been altered to ferrihydrite-dominated assemblages. b) Granular pyroxene and olivine type 1 CgMM with altered FeNi metal. c) Porphyritic olivine type 2 CgMM with an interstitial Na-bearing aluminosilicate glass. d) Porphyritic olivine and pyroxene type 2 CgMM with interstitial glassy mesostasis. This particle also contains some diopside. e) Radiate pyroxene type 2 particle with small iron-oxide dendrites in the interstitial glass. f) A single crystal CgMM of diopside. This particle lacks a magnetite rim (the bright exterior is charging), however, its extraterrestrial nature is indicated by the FeNi metal droplets it contains. Note that particles c and d both exhibit igneous rims similar to those of ScMM and unmelted MMs (cf. Genge 2005). Scale bar is 50 μm.

Minor Phases in FgMMs

In addition to isolated pyroxene and olivine, minor primary phases found in FgMMs include magnetite, tochilinite, metal, sulfide, and refractory minerals typical of CAIs. These presence of these phases have some implications for the parent body associations of micrometeorites but are not usually strongly discriminative. Magnetite framboids and platelets, for example, although most common in CI1 chondrites are also found in Tagish Lake, CM2 and CR2 chondrites, and more rarely in CV3 and CO3 chondrites. Tochilinite, or at least its thermal decomposition products, nevertheless, may be
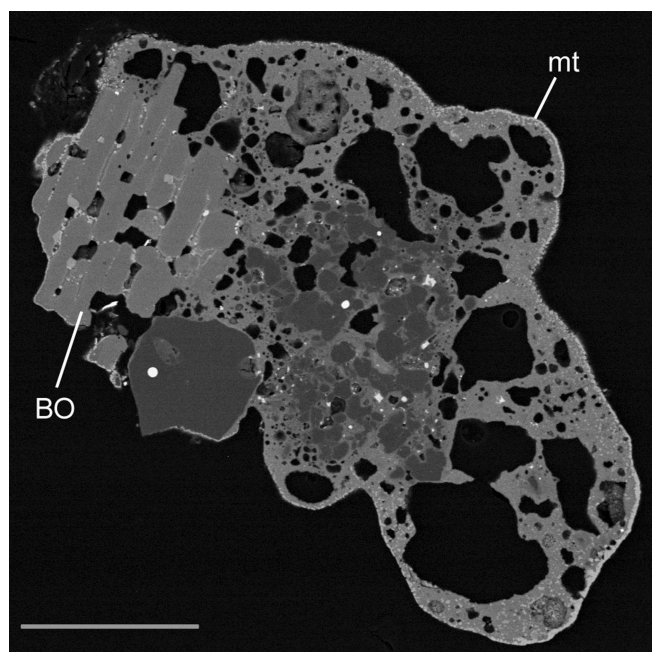
Fig. 6. Backscattered electron image of a Sc/CgMM from the South Pole water well containing a fragment of a barred olivine igneous object (BO) covering the upper quarter of the particle. The particle has a well developed igneous rim. Scale bar is 50 µm.

diagnostic of a CM2 affinity (Brearley et al. 1998). The presence of minor phases in FgMMs is probably, therefore, best recorded as a prefix giving classifications, such as, "framboidal magnetite-bearing FgMM" or "spinel-bearing FgMM."

A special case of minor phases in FgMMs, however, are those that contain refractory phases, such as Mg-Al-spinel, surrounded by iron-rich phyllosilicates. These particles are strongly reminiscent of aqueously altered CAIs from CM2 chondrites and are discussed in more detail below.

Unmelted MMs: Coarse-Grained MMs

The majority of CgMMs have igneous textures and are dominated by pyroxene and/or olivine within a glassy mesostasis which can contain accessory metal, sulfide, and/or iron oxides. The majority of these particles have broadly chondritic mineral assemblages similar to chondrules or the primitive achondrites. Most other CgMMs are fragments of single crystals of olivine and pyroxene and thus provide little genetic information to suggest an adequate parent body association. Very rare achondritic particles, with textures and mineralogies suggesting derivation from differentiated parent have been recognized such as a single particle containing pigeonite with augite exsolution and anorthite that may have an affinity to basaltic achondrites (Gounelle et al. 2005a). Consequently CgMMs should probably be subdivided into (a) chondritic CgMMs, which have broadly chondritic mineralogies (Figs. 5a–e), (b) rare achondritic CgMMs, which have non-chondritic, non-refractory mineralogies, and (c) single crystal

CgMMs (Fig. 5f). Under a binocular microscope CgMMs can sometimes be identified by their facetted surfaces, however, they are often difficult to distinguish from ScMMs, particularly when they have igneous rims.

Chondritic CgMMs

Chondritic CgMMs can be subdivided using a petrological scheme (Genge et al. 2005) in which particles are classified in a similar fashion to chondrules. Coarse-grained MMs can, therefore, be porphyritic (P), granular (G) or radiating pyroxene (Rp). Porphyritic and granular particles, like chondrules, are prefixed with their dominant mineralogies following their textural classification, for example, into porphyritic olivine (Po), porphyritic pyroxene (PP), porphyritic olivine and pyroxene (POP) (Fig. 5). A single Sc/CgMM containing an area of barred olivine similar to barred olivine chondrules has been identified (S. Taylor, unpublished data; Fig. 6), the term barred olivine (Bo) should be used for such particles, when dominated by the coarse-grained portion.

Chondritic, porphyritic, and granular CgMMs fall into two main chemical populations based on the compositions of their olivines and pyroxenes, and the abundance of metal and sulfide (Genge 2006). This scheme is analogous to that of chondrules. Type 1 CgMMs are those with Mg-rich olivines and/or pyroxenes, and usually contain abundant FeNi metal and sulfides (Figs. 5a and 5b). Type 2 CgMMs are those with more Fe-rich olivines and pyroxenes, less abundant metal and sulfides, and sometimes contain iron oxides (Figs. 5c, 5d, and 5e). The boundary between the two groups, as with chondrules, is 10 mol% fayalite or ferrosilite (Grossman et al. 1988). Coarse-grained MMs consisting of single crystals are also observed (Fig. 5f).

In classifying primitive CgMMs into type 1 and type 2 particles it should not be assumed that these necessarily originated as fragments of chondrules. A derivation from a primitive achondrite source or even shock melts is possible for many individual particles.

Chondritic CgMMs sometimes include selvages of fine-grained matrix similar to that of FgMMs (Fig. 7). These particles indicate that at least a proportion of CgMMs and FgMMs are derived from the same parent bodies and have been termed composite MMs (Genge 2006). Such particles are classified as Fg/CgMMs.

Chondritic CgMMs with subspherical particle shapes (Fig. 8) are difficult to distinguish from cosmic spherules since both can have similar igneous textures formed by crystallization during rapid cooling. Chondritic CgMMs, however, can be identified on the basis of any of the following criteria: (1) the presence of pyroxene phenocrysts, which are absent within cosmic spherules since pyroxene crystallization is inhibited by the extremely rapid cooling rates (several hundred °C s^{-1}) experienced by spherules (Taylor and Brownlee 1991), (2) the occurrence of a magnetite rim, which

is rare on cosmic spherules and usually associated with metal separation, and (3) the presence of an igneous rim similar to that found on some fine-grained micrometeorites. It should be noted, however, that pyroxene dendrites can occasionally be found in cosmic spherules within the mesostasis and are probably formed at very large supercoolings (Genge et al. 1997a).

Unmelted MMs: Refractory MMs

Micrometeorites that potentially represent samples of components other than chondrules (i.e., other than chondritic igneous objects) and matrix have been reported but are present in low abundances within MM collections such that sufficient numbers have yet to be characterized to facilitate a reliable classification scheme. Principal among these are unmelted FgMM containing refractory minerals that are likely to be fragments of refractory inclusions. Fine-grained micrometeorites containing isolated grains of spinel, perovskite, melilite, fassaite, and hibonite have been reported by previous studies (Beckerling and Bischoff 1995; Engrand and Maurette 1998; Engrand 1999; Gounelle 2000; Greshake et al. 1996; Hoppe et al. 1995; Kurat et al. 1994; Lindstrom et al. 1992) and some have been classified on the basis of their trace element systematics following the scheme used for CAIs (Greshake et al. 1996). In the current system it is recommended that such particles are classified as FgMMs and prefixed with the refractory mineral name (e.g., hibonite-bearing FgMM) since the refractory minerals comprise a volumetrically small component of the particles.

Several micrometeorites, however, have been discovered that are dominated by refractory minerals, or their alteration products, that warrant a separate classification. Particles containing spinels surrounded by Fe-rich phyllosilicates (or their thermal decomposition products) bear a close resemblance to aqueous-altered CAIs from CM2 chondrites (Fig. 10a) (Gounelle 2000; Kurat et al. 1994). These particles are classified as hydrated refractory MMs.

Only three micrometeorite (BI54B3-31, 94-4-36, and 94-19-5) dominated by refractory minerals have been previously reported. B154B3-31 is a porous aggregate of anhedral fassaite with a smaller region dominated by enstatite (Greshake et al. 1996). Particles 94-4-36 and 94-19-5 are dominated by spinel containing inclusions of perovskite and surrounded by Fe-rich phyllosilicates in the case of 94-4-5 (Fig. 10c) and by Ca-rich pyroxene and matrix in the case of 94-4-36 (Fig. 10b) (Engrand et al. 1995). Figure 10d shows a fourth refractory particle dominated by aluminous diopside with minor forsterite (Genge, unpublished data). Classifying these refractory MMs is problematic due to the low numbers of particles so far discovered and the lack of a unified petrological CAI classification system. It is suggested, therefore, that refractory particles should simply be
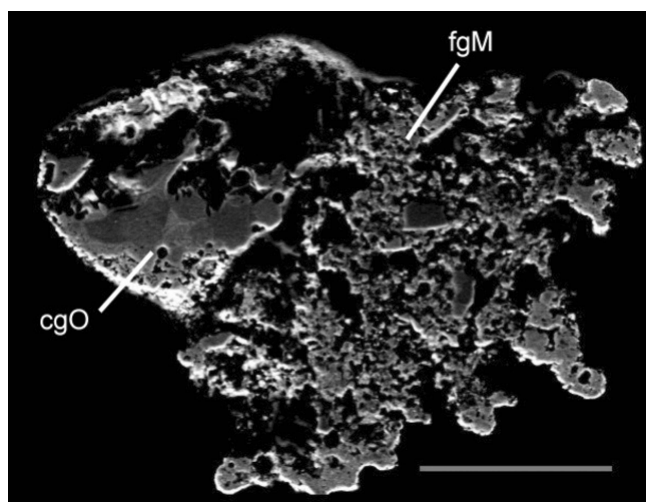

Fig. 7. A backscattered electron image of a Fg/CgMM micrometeorite (SP94-100-001) consisting of both fine-grained (FGM) and coarse-grained (IO) portions. Scale bar is 50 μm.

subdivided into porous and compact varieties and prefixed with their dominant mineralogy. Particle BI54B3-31 would, therefore, be a fassaite-enstatite-bearing, porous refractory MM. The low abundance of CAI-related MMs, although similar to CM2 abundances, may be a result of separation techniques used in the laboratory, which focus on dark or spherical particles. The mineralogy and grain-sizes of refractory MMs are, however, subtly different from those of CM2 chondrites (Brearley and Jones 1998).

Unmelted MMs: Ultracarbonaceous MMs

Three large particles (>200 μm) have now been identified amongst those collected from Antarctic snow that contain considerably higher abundances of carbon than CI chondrites (Nakamura et al. 2005). The MMs are dominated by networks of amorphous carbon containing olivine, low-Ca pyroxene, pyrrhotite, and kamacite. The silicate components of these MMs have heterogeneous compositions that suggest they are unequilibrated. One of the particles, KWP3F5, contained melted silicates suggesting intense atmospheric heating, however, another contained high abundances of presolar silicates and had evidently survived atmospheric entry without significant heating. These MMs may be cometary in origin and are termed ultracarbonaceous MMs by their discoverers.

DISCUSSION

A discussion of the more controversial aspects of micrometeorite origins and their implications is not appropriate within the current work which aims to provide a balanced framework for particle classification based on features likely to be of genetic importance. There are,
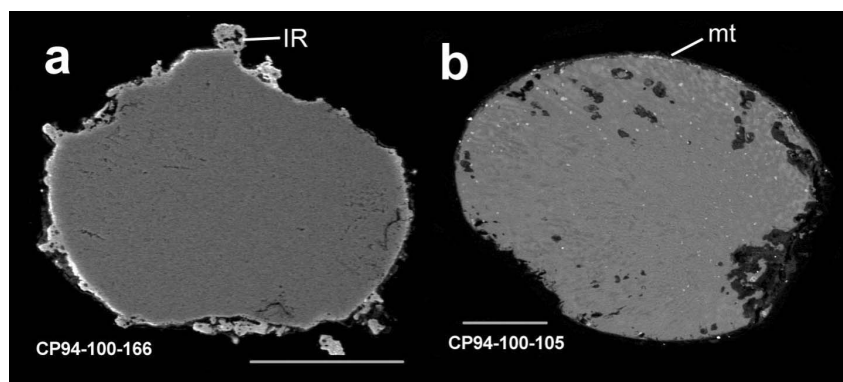
Fig. 8. Backscattered electron images of subspherical CgMMs. a) A radiating pyroxene type 2 CgMM. This particle has a vesicular igneous rim (IR) consisting fayalitic olivine microphenocrysts with interstitial glass. b) A radiating pyroxene type 2 CgMM with interstitial olivine and glass. This particle has a magnetite rim (mt). Scale bar is 50 μm.
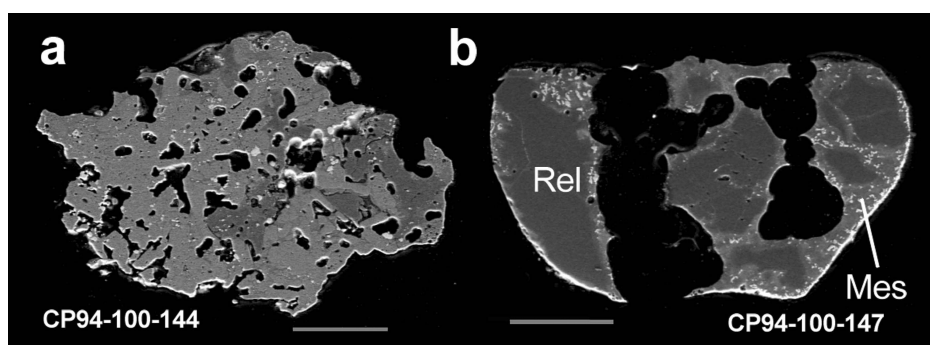

Fig. 9. Atmospheric heating effects in CgMMs. a) A vesicular CgMM in which vesicles have formed within glassy mesostasis. Note Fig. 5b, 5d, and 5e are also vesicular particles. b) A partially melted CgMM containing relict anhydrous silicates (Rel) within melted mesostasis (Mes). Although having experienced a significant degree of melting this particle retains an irregular shape. Coarse-grained spherules are shown in Figs. 1x and 1y. Scale bar is 50 μm.

however, a number of outstanding problems relating to the genetic interpretation of particles that are worthy of mention here. In particular these relate to particle types that we might expect to find among micrometeorite collections that have yet to be recognized since potentially these will require the current classification system to be extended.

In the following we discuss particle types we might expect to be present within micrometeorite collections based on the assumptions that: (1) both asteroidal and cometary dust 30–1000 μm contribute to collections retrieved from the Earth's surface, (2) asteroidal dust will include particles derived from asteroids with similar components to both chondritic and achondritic meteorites, and (3) cometary dust, at least prior to atmospheric entry, will probably include highly primitive materials similar to anhydrous IDPs.

There is, however, little agreement, even amongst the coauthors of this paper, on the exact distinction between cometary and asteroidal materials. It has often been assumed, for example, that liquid water cannot exist on cometary nuclei and thus cometary materials will not include phyllosilicates. The recent preliminary detection of hydrous silicates in

comet Temple 1 by the Deep Impact mission (unpublished press release), however, indicates that some FgMMs and even CI chondrites may have a cometary origin (Gounelle et al. 2004, 2006; Lodders and Osborne 1999; McSween and Weissman 1989). The occurrence of chondrules and refractory inclusions within cometary materials has also been postulated to explain rare dense compact particles observed within cometary meteor streams (Swindle and Campins 2004) and is not precluded by models of the transport of these objects in the early solar system such as the X-wind model (Shu et al. 2001). The presence of chondrule or CAI fragments in MMs, therefore, is not conclusive evidence of an asteroidal origin. Similarly ultracarbonaceous MMs and anhydrous IDPs are assumed to be derived from comets on the basis of the presence of abundant presolar silicates and their high C-contents (Bradley 1994; Nakamura et al. 2005), however, these might also be derived from highly primitive objects in the main asteroid belt. The additional particle types, discussed below, therefore, are where possible presented without reference to parent body type since it is entirely possible that cometary and primitive asteroidal matter are closely related.
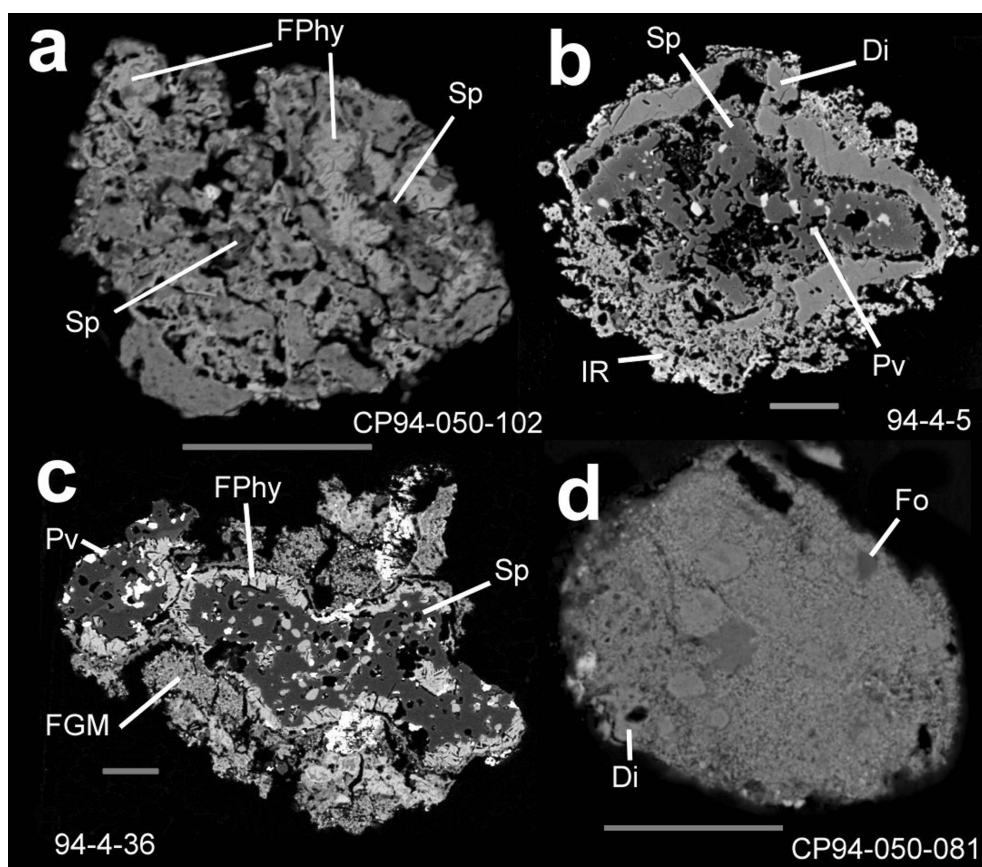
Fig. 10. Backscattered electron images of CgMMs. a) A hydrated refractory MM dominated by coarse-grained Fe-rich phyllosilicates (FPhy) and containing spinel (Sp, $MgAl_2O_4$). b) A compact spinel-diopside (Di)–perovskite (Pv) refractory MM with a scoriaceous igneous rim. c) A hydrated spinel-perovskite refractory MM with fine-grained matrix (FGM). d) A porous diopside-forsterite (Fo) refractory MM. Scale bar is 50 µm.

Achondritic Micrometeorites

Currently only a single particle has been reported that is demonstrably achondritic since it contains a large pigeonite crystal with augite exsolution lamellae and so demonstrably experienced slow cooling. This particle, which contains plagioclase, also has a depleted differentiated REE pattern suggesting crystallization from a melt generated from a depleted lithosphere (Gounelle et al. 2005a).

Several problems exist in the identification of achondritic particles, of which grain-size is probably the most significant. The large grain-sizes of most achondrites are likely to mean that dust derived from stony achondritic sources are present as single crystals, primarily of olivine or pyroxene that are unlikely to be distinct from those of primitive meteorites. Identifying an achondritic source for single crystal particles would involve trace element analyses and/or isotopic data. Consequently surveys of large numbers of particles are not practical. Only those phases with distinctive major element chemistries, such as kirschsteinite (angrites) and augite (basaltic achondrites), may be readily identified. Exsolution, in particular within pyroxenes, which indicates slow cooling

may, however, also provide an indicator of an achondritic origin by which particles may be selected for more detailed chemical or isotopic analysis. Materials from primitive achondrites are likely to be particularly problematic due to their broadly chondritic mineral assemblages and trace element systematics (Mittlefehdt et al. 1998). Martian or lunar MMs that might be expected to occur within dust collections would be classified as achondritic MMs and their identification would rely on their petrology and isotopic analysis. Martian MMs are likely to be exceedingly rare since martian interplanetary dust particles will inherently evolve to Earth-crossing orbits on short timescales (Dohnanyi 1976) and they can only be produced by relatively recent impacts on Mars. In contrast, lunar MMs are likely to contribute significantly to the terrestrial dust flux for short periods following impact events on the Moon.

Entirely metallic MMs, representing dust-sized debris from metallic asteroids, or primitive metal from chondrites, have also yet to be recognized. Isotopic analyses of iron-type spherules (Engrand et al. 1998, 2005; Herzog et al. 1999), however, suggest that many of these particles may have been metallic FeNi micrometeoroids that have been oxidized

during atmospheric entry heating. The absence of unmelted metallic MMs, however, is puzzling since photometry studies of M-type asteroids imply the presence of regolith (Belskaya and Lagerkrist 1996). The presence of metal within CgMMs certainly suggests it can survive long term storage in Antarctic ice without complete oxidation. A density bias during collection of melt water is a possible explanation, however, collections derived from melting of Antarctic snow have also not yet reported metallic unmelted MMs (Duprat et al. 2003; Nakamura et al. 1999).

Chondrite-Like Micrometeorites

There are a significant number of unresolved issues related to particles derived from chondrite-like parent bodies. Materials similar to components within the known chondritic meteorite groups would be expected to occur within MM collections, however, to date most micrometeorites appear to have affinities with C1 and C2 parent bodies. These also appear to sample mainly matrix, isolated minerals, and chondrules, with rare materials from CAIs. It is likely that the absence of MMs with affinities to other chondrite groups and components of chondrites relate to either a lower abundance within the asteroid belt or selection biases due to dust production, capture by the Earth, or survival of atmospheric entry.

The absence of FgMMs with affinities to CV3 or CO3 matrix is particularly surprising considering the significant number of meteorites of these classes. One possible reason for this under-representation may be if the Eos family of main belt asteroids are the parent bodies of C3 meteorites, as suggested by asteroid spectroscopy (Nesvorny et al. 2003), since the dust band associated with this asteroid has a relatively high inclination. The latitude of CV3 and CO3 dust at 1 AU may, therefore, result in less of this material being captured by the Earth than from other, lower inclination sources (Kortenkamp and Dermett 1999). CV3 and CO3 dust may be present in low abundances in MM collections, however, particles dominated by porous assemblages of relatively fayalitic olivine, that might be analogous to CV3 and CO3 matrix (Brearley and Jones 1998), have yet to be identified.

Particles with affinities to reduced, primitive parent bodies such as the enstatite chondrites, and from highly oxidized parent bodies, such as R-chondrites, have also yet to be discovered. One type I Po CgMM, containing Si-bearing metal and silica, could have affinities to the K-chondrites or enstatite chondrites (Genge et al. 1997b), and several type II CgMMs with highly fayalitic olivines (e.g., Fa_{48}) have been identified (Genge, unpublished data) that could be derived from oxidized chondrite groups. Currently the low abundance of such particles does not warrant extension of the classification scheme for CgMMs.

Classifying CgMMs derived as chondrule fragments from parent bodies with affinities to ordinary chondrites is a particular problem since without the additional information on chondrule abundances and bulk chemistry provided by macroscopic samples these materials will be difficult to distinguish as individual particles from those derived from carbonaceous chondrites. It has been suggested that CgMMs could potentially be classified into equilibrated or unequilibrated particles, relating to type 1–3 and type 4–7 chondrites, on the basis of textural, mineralogical and chemical properties (Genge and Grady 1998b). Coarse-grained MMs containing glassy mesostases, having chemical zoning in their phenocrysts, are demonstrably unequilibrated, although such features must be shown not to have arisen during atmospheric entry. Identifying equilibrated particles is more problematic since lack of chemical zoning or glassy mesostasis does not necessarily indicate equilibration. FeO, MnO, CaO, and Cr_2O_3 abundances within olivines may, however, provide one diagnostic test (Brearley and Jones 1998).

Several C2 FgMMs have been reported in which the fine-grained matrix had spatial chemical variations of Fe/Mg closely resembling relict hydrated chondrules found in CM2 chondrites (Genge 2002). Whether such particles should be given a classification denoting such an origin is debatable since recognizing such particles is dependent on preservation of their pseudomorphed textures and it is, therefore, difficult to give quantitative criteria for their identification.

One further possibility must be considered for those particles classified as altered FgMMs that relates directly to the nature of their parent body. Implicit in the classification of a FgMM as altered is that the thermal alteration occurred during atmospheric entry. Another possibility, however, is that some altered FgMMs may represent hydrous matrix that was dehydrated due to parent body metamorphism, similar to the suggested origin of dark clasts found in a range of meteorites (Brearley and Jones 1998). Distinguishing between the effects of parent body and atmospheric heating is likely to involve detailed characterization of textures and mineralogies at the submicron scale and thus would be difficult to apply in particle surveys. The complete absence of a magnetite rim would, perhaps, be one criteria that could be used to indicate minimal entry heating. Conversely, those particles with igneous rims, which are supported by the exothermic dehydration reactions, demonstrably contained hydrated silicates during atmospheric entry (Genge 2006).

Shock State

No studies have yet examined the shock state of MMs, mainly because standard thin-section techniques for the identification of shock features in particular phases are difficult to apply to small particulates. Several textural features, which can be characterized in backscattered electron images, may, however, prove to be useful in characterizing shock in MMs. Metal-sulfide veining within anhydrous silicates is a common shock feature within chondritic

meteorites that can be readily identified, but gives little correlation with the shock scale used for meteorites (Stoeffler et al. 1991). Mosaicism within olivine may provide a means of identifying particles shocked to S4 and above, however, discriminating mosaicism from annealed granular textures in BEI images is likely to be inconclusive. Both metal-sulfide veining and mosaicism may, nevertheless, provide criteria by which particles can be selected for TEM characterization of microstructural defects to better constrain their shock state. Annealing during atmospheric entry will also complicate analysis of shock features.

Micrometeorites sampling shock melts are also likely to be difficult to distinguish from fragments of chondrules and primitive achondrites since these have broadly similar mineralogies and textures. The presence of abundant submicron FeNi metal blebs and annealed relicts may, however, provide criteria for their identification.

Highly Primitive Micrometeorites

Models of atmospheric entry heating suggest that at least a proportion of low geocentric velocity cometary dust, which would be expected to be highly primitive dust, should survive atmospheric entry at diameters up to ~100 μm (Flynn 1991; Genge 2003; Liou and Zook 1996; Love and Brownlee 1991). Most of these cometary MMs are predicted to experience significant thermal alteration, however, a small proportion are expected to survive atmospheric entry without melting. To date only the ultracarbonaceous MMs have been identified as such highly primitive, potentially cometary, micrometeorites.

One possible explanation for the paucity of MMs equivalent to anhydrous IDPs may be that these particles are extremely fragile and become disaggregated during atmospheric flight, storage in Antarctic ice and/or collection. Certainly preliminary results of a new collection made by melting of Antarctic snow indicate the presence of larger numbers of fluffy, fragile particles (Duprat et al. 2003, 2005). An alternative explanation is that cometary particles are present among MMs but more closely resemble asteroidal materials than has previously been thought.

Intensely heated cometary particles have also not been identified with any certainty. An unusual Mg-rich ScMM containing segregated FeNi metal droplets might represent one such particle on the basis of the carbon-enriched nature of anhydrous IDPs and the significant reduction of the particle (Genge and Grady 1998c). No other similar particles, however, have yet been discovered.

CONCLUSIONS

The field of "micrometeoritics" has progressed considerably in the last two decades since the pioneering collection of particles from the deep sea (Brownlee 1985)

and Antarctic ice (Maurette et al. 1991). The characterization of many thousands of MMs by authors of this paper and others has now allowed groups and classes of particle to be identified and placed into a genetically significant framework. In the future, as the work of characterization of these materials progresses, rare types of MM will undoubtedly be recognized and allow MMs to be used to better constraint the diversity of asteroidal and cometary bodies within our solar system.

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REFERENCES

Alexander C. M. O'D. 2002. Mass-dependent fractionation of Mg, Si, and Fe isotopes in five stony cosmic spherules. *Geochimica et Cosmochimica Acta* 66:173–183.

Beckerling W. and Bischoff A. 1995. Occurrence and composition of relict minerals in micrometeorites from Greenland and Antarctica—Implications for their origins. *Planetary and Space Science* 43:435–449.

Belskaya I. N. and Lagerkvist C. I. 1996. Physical properties of M class asteroids. *Planetary and Space Science* 44:783–794.

Blanchard M. B., Brownlee D. E., Bunch T. E., Hodge P. W., and Kyte F. T. 1980. Meteoroid ablation spheres from deep sea sediments. *Earth and Planetary Science Letters* 46:178–190.

Bradley J. P. 1994. Chemically anomalous, preaccretionally irradiated grains in interplanetary dust from comets. *Science* 265: 925–929.

Brearley A. J. and Jones R. H. 1998. Chondritic meteorites. In *Planetary materials*, vol. 36, edited by Papike J. J. Washington, D.C.: Mineralogical Society of America. pp. 1–191.

Brownlee D. E. and Bates B. 1983. Meteor ablation spherules as chondrule analogs. In *Chondrules and their origins*, edited by King E. A. Houston: Lunar and Planetary Institute. pp. 10–25.

Brownlee D. E. 1985. Cosmic dust: Collection and research. *Annual Reviews of Earth and Planetary Sciences* 13:147–173.

Brownlee D. E., Love S., and Schramm L. S. 1991. Cosmic spherules and giant micrometeorites as samples of main belt asteroids (abstract). 22nd Lunar and Planetary Science Conference. p. 147.

Brownlee D. E., Bates B., and Schramm L. 1997. The elemental composition of stony cosmic spherules. *Meteoritics & Planetary Science* 32:157–175.

Clayton R. N., Mayeda T. M., and Brownlee D. E. 1986. Oxygen isotopes in deep-sea spherules. *Earth and Planetary Science Letters* 79:235–240.

Dohnanyi J. S. 1967. Sources of interplanetary dust: Asteroids. In

Interplanetary dust and the zodiacal light, vol. 48, edited by Elsasser H. and Fechtig H. Berlin: Springer-Verlag. pp. 187–206.

Dohnanyi J. S. 1976. Sources of interplanetary dust: Asteroids. In *Interplanetary dust and zodiacal light*, edited by Elsasser H. Berlin: Springer-Verlag. pp. 569–639.

Duprat J., Engrand C., Maurette M., Gounelle M., Hammer C., and Kurat G. 2003. The CONCORDIA collection: Pristine contemporary micrometeorites from central Antarctica surface snow (abstract #1727). 34th Lunar and Planetary Science Conference. CD-ROM.

Duprat J., Engrand C., Maurette M., Gounelle M., Kurat G., and Leroux H. 2005. Friable micrometeorites from central Antarctic snow (abstract #1678). 36th Lunar and Planetary Science Conference. CD-ROM.

Engrand C., Maurette M., Zolensky M. E., Kurat G., and Walter J. 1995. Electron microprobe analyses of Antarctic micrometeorites and interplanetary dust particles collected in the stratosphere (abstract). 26th Lunar and Planetary Science Conference. p. 375.

Engrand C. and Maurette M. 1998. Carbonaceous micrometeorites from Antarctica. *Meteoritics & Planetary Science* 33:565–580.

Engrand C., McKeegan K. D., Leshin L. A., and Brownlee D. E. 1998b. In situ measurement of oxygen isotopic compositions of deep-sea and Antarctic cosmic spherules (abstract #1473). 29th Lunar and Planetary Science Conference. CD-ROM.

Engrand C. 1999. Oxygen isotopic compositions of individual minerals in Antarctic micrometeorites: Further links to carbonaceous chondrites. *Geochimica et Cosmochimica Acta* 63:2623–2636.

Engrand C., Deloule E., Robert F., Maurette M., and Kurat G. 1999. Extraterrestrial water in micrometeorites and cosmic spherules from Antarctica: An ion microprobe study. *Meteoritics & Planetary Science* 34:773–786.

Engrand C., McKeegan K. D., Leshin L. A., Herzog G. F., Schnabel C., Nyquist L. E., and Brownlee D. E. 2005. Isotopic compositions oxygen, iron, chromium, and nickel in cosmic spherules: Toward a better comprehension of atmospheric entry heating effects. *Geochimica et Cosmochimica Acta* 69:5365–5385.

Feng H., Jones K. W., Tomov S., Stewart B., Herzog G. F., Schnabel C., and Brownlee D. E. 2005. Internal structure of type I deep-sea spherules by X-ray computed microtomography. *Meteoritics & Planetary Science* 40:195–206.

Fisher G. L., Chang D. P. Y., and Brummer M. 1976. Fly ash collected from electrostatic precipitators: Microcrystalline structures and the mystery of the spheres. *Science* 7:553–555.

Flynn G. J. 1991. Survival of large micrometeorites on atmospheric entry: Implications for their sources and the flux of cometary dust (abstract). 22nd Lunar and Planetary Science Conference. p. 393.

Flynn G. J. 1995. Thermal gradients in interplanetary dust particles: The effect of an endothermic phase transition (abstract). 26th Lunar and Planetary Science Conference. p. 405.

Genge M. J., Hutchison R., and Grady M. M. 1996. Atmospheric alteration of coarse-grained Antarctic micrometeorites (abstract). *Meteoritics & Planetary Science* 31:A49.

Genge M. J., Grady M. M., and Hutchison R. 1997a. The textures and compositions of fine-grained Antarctic micrometeorites— Implications for comparisons with meteorites. *Geochimica et Cosmochimica Acta* 61:5149–5162.

Genge M. J., Grady M. M., and Hutchison R. 1997b. Free-silica in a non-chondritic micrometeorite from Antarctic ice (abstract). 28th Lunar and Planetary Science Conference. p. 405.

Genge M. J. and Grady M. M. 1998a. Melted micrometeorites from Antarctic ice with evidence for the separation of immiscible Fe-Ni-S liquids during entry heating. *Meteoritics & Planetary Science* 33:425–434.

Genge M. J. and Grady M. M. 1998b. A petrological-chemical classification scheme for coarse-grained micrometeorites. *Meteoritics & Planetary Science* 33:A56.

Genge M. J. and Grady M. M. 1998c. Melted micrometeorites from Antarctic ice with evidence for the separation of immiscible Fe-Ni-S liquids during entry heating. *Meteoritics & Planetary Science* 33:425–434.

Genge M. J. and Grady M. M. 1999. The fusion crusts of stony meteorites: Implications for the atmospheric reprocessing of extraterrestrial materials. *Meteoritics & Planetary Science* 34:341–356.

Genge M. J., Engrand C., and Grady M. M. 2000. The parent bodies of thermally altered fine-grained micrometeorites: Comparisons with CI and CM fusion crusts (abstract). *Meteoritics & Planetary Science* 35:A59.

Genge M. J., Bradley J. P., Engrand C., Gounelle M., Harvey R. P., and Grady M. M. 2001. The petrology of fine-grained micrometeorites: Evidence for the diversity of primitive asteroids (abstract #1546). 32nd Lunar and Planetary Science Conference. CD-ROM.

Genge M. J. 2002. Hydrated chondrule fragments among micrometeorites (abstract). *Meteoritics & Planetary Science* 37:A51.

Genge M. J. and Grady M. M. 2002. The distribution of asteroids: Evidence from Antarctic micrometeorites (abstract #1010). 33rd Lunar and Planetary Science Conference. CD-ROM.

Genge M. J. 2003. Primary variations in micrometeorites with entry velocity (abstract #1151). 34th Lunar and Planetary Science Conference. CD-ROM.

Genge M. J., Gileski A., and Grady M. M. 2005. Chondrules in Antarctic micrometeorites. *Meteoritics & Planetary Science* 40:225–238.

Genge M. J. 2006. Igneous rims on micrometeorites. *Geochimica et Cosmochimica Acta* 70:2603–2621.

Glass B. P. 1990. Tektites and microtektites: Key facts and interferences. *Tectonophysics* 171:393–404.

Gounelle M. 2000. Matière extraterrestre sur Terre: Des Océans aux protoétoiles. Ph.D. thesis, Universite Paris. In French.

Gounelle M., Devouard B., Engrand C., Genge M. J., Toppani A., and Leroux H. 2002. TEM study of Antarctic micrometeorites: A preliminary report (abstract). *Meteoritics & Planetary Science* 37:A55.

Gounelle M., Spurný P., and Bland P. 2004. The orbit of the Orgueil meteorite from historical records (abstract). *Meteoritics & Planetary Science* 39:A45.

Gounelle M., Engrand C., Chaussidon M., Zolenksy M. E., and Maurette M. 2005a. An achondritic micrometeorite from Antarctica: Expanding the solar system inventory of basaltic asteroids (abstract #1655). 36th Lunar and Planetary Science Conference. CD-ROM.

Gounelle M., Engrand C., Maurette M., Kurat G., McKeegan K. D., and Brandstaetter F. 2005b. Small Antarctic micrometeorites: A mineralogical and in site mineralogical study. *Meteoritics & Planetary Science* 40:917–932.

Gounelle M., Spurný P., and Bland P. A. 2006. The orbit and atmospheric trajectory of the Orgueil meteorite from historical records. *Meteoritics & Planetary Science* 41:135–150.

Greshake A., Hoppe P., and Bischoff A. 1996. Mineralogy, chemistry, and oxygen isotopes of refractory inclusions from stratospheric interplanetary dust particles and micrometeorites. *Meteoritics* 31:739–748.

Greshake A., Kloeck W., Arndt P., Maetz M., Flynn G. J., Bajt S., and Bischoff A. 1998. Heating experiments simulating atmospheric entry heating of micrometeorites: Clues to their parent body sources. *Meteoritics & Planetary Science* 33:267–290.

Grossman J. N., Rubin A. E., Nagahara H., and King E. A. 1988.

Properties of chondrules. In *Meteorites and the early solar system*, edited by Kerridge J. F. and Matthews M. S. Tucson, Arizona: The University of Arizona Press. pp. 619–659.

Harvey R. P., Dunbar N. W., McIntosh W. C., Esser R. P., Nishiizumi K., Taylor S., and Caffee M. W. 1998. Meteoritic event recorded in Antarctic ice. *Geology* 26:607–610.

Herzog G. F., Xue S., Hall G. S., Nyquist L. E., Shih C. Y., Weismann H., and Brownlee D. E. 1999. Isotopic and elemental composition of iron, nickel, and chromium in type I deep-sea spherules: Implications for origin and composition of the parent micrometeoroids. *Geochimica et Cosmochimica Acta* 63:1443–1457.

Hoppe P., Kurat G., Walter J., and Maurette M. 1995. Trace elements and oxygen isotopes in a CAI-bearing micrometeorite from Antarctica (abstract). 26th Lunar and Planetary Science Conference. p. 623.

Iwata N. and Imae N. 2001. The collection of Antarctic micrometeorites by JARE-41 in 2000 (abstract). *Meteoritics & Planetary Science* 36:A89.

Kortenkamp S. J. and Dermott S. F. 1999. Accretion of interplanetary dust particles by the Earth. *Icarus* 135:469–495.

Kurat G., Koeberl C., Presper T., Franz B., and Maurette M. 1994. Petrology and geochemistry of Antarctic micrometeorites. *Geochimica et Cosmochimica Acta* 58:3879–3904.

Lindstrom D. J. and Klock E. 1992. Analyses of 24 unmelted Antarctic micrometeorites by instrumental neutron activation analysis. *Meteoritics* 27:250.

Liou J. C. and Zook H. A. 1996. Comets as a source of low eccentricity and low inclination interplanetary dust particles. *Icarus* 123:491–502.

Lodders K. and Osborne R. 1999. Perspectives on the comet-asteroid-meteorite link. *Space Science Reviews* 90:289–297.

Love S. G. and Brownlee D. E. 1991. Heating and thermal transformation of micrometeoroids entering the Earth's atmosphere. *Icarus* 89:26–43.

Maurette M., Olinger C., Michel-Levy M. C., Kurat G., Pourchet M., Brandstatter F., and Bourot-Denise M. 1991. A collection of diverse micrometeorites recovered from 100 tonnes of Antarctic blue ice. *Nature* 351:44–47.

McSween H. Y. Jr. and Weissman P. R. 1989. Cosmochemical implications of the physical processing of cometary nuclei. *Geochimica et Cosmochimica Acta* 53:3263–3271.

Mittlefehdt D. W., McCoy T. J., Goodrich C. A., and Kracher A. 1998. Non-chondritic meteorites from asteroidal bodies. In *Planetary materials*, vol. 36, edited by Papike J. J. Washington, D.C.: Mineralogical Society of America. pp. 4-1–4-195.

Nakamura K., Nogushi T., Ozono Y., Osawa T., and Nagao K. 2005. Mineralogy of ultracarbonaceous large micrometeorites (abstract). *Meteoritics & Planetary Science* 40:A110.

Nakamura T., Imae N., Nakai I., Noguchi T., Yano H., Terada K., Murakami T., Fukuoka T., Nogami K.-I., Ohashi H., Nozaki W., Hashimoto M., Kondo N., Matsuzaki H., Ichikawa O., and Ohmori R. 1999. Antarctic micrometeorites collected at the Dome Fuji Station. 23rd Symposium on Antarctic Meteorites. p. 183.

Nakamura T., Noguchi T., Yada T., Nakamuta Y., and Takaoka N. 2001. Bulk mineralogy of individual micrometeorites determined by X-ray diffraction analysis and transmission electron microscopy. *Geochimica et Cosmochimica Acta* 65:4385–4397.

Nesvorny D., Bottke W. F., Levison H. F., and Dones L. 2003. Recent origin of the solar system dust bands. *The Astrophysical Journal* 591:486–497.

Nininger H. H. 1956. *Arizona's Meteor Crater.* Sedona, Arizona: World Press. 232 p.

Noguchi T. and Nakamura T. 2000. Mineralogy of Antarctic micrometeorites recovered from the Dome Fuji Station. 24th Symposium on Antarctic Meteorites, p. 285.

Noguchi T., Nakamura T., and Nozaki W. 2002. Mineralogy of phyllosilicate-rich micrometeorites and comparison with Tagish Lake and Sayama meteorites. *Earth and Planetary Science Letters* 202:229–246.

Olinger C. T. 1990. Isotopic measurements of solar noble gases in individual micrometeorites from Greenland and Antarctica. Ph.D. thesis, University of Washington.

Raisbeck G. M. and Yiou F. 1987. ^{10}Be and ^{26}Al in micrometeorites from Greenland ice. *Meteoritics* 22:485–486.

Raisbeck G. M. and Yiou F. 1989. Cosmic-ray exposure ages of cosmic spherules (abstract). *Meteoritics* 24:318.

Rietmeijer F. J. M. 1998. Interplanetary dust particles. In *Planetary materials*, vol. 36, edited by Papike J. J. Washington, D. C.: Mineralogical Society of America. pp. 28–119.

Shu F. H., Shang H., Gounelle M., Glassgold A. E., and Lee T. 2001. The origin of chondrules and refractory inclusions in chondritic meteorites. *The Astrophysical Journal* 548:1029–1050.

Stöffler D., Keil K., and Scott E. R. D. 1991. Shock metamorphism of ordinary chondrites. *Geochimica et Cosmochimica Acta* 55:3845–3867.

Swindle T. D. and Campins H. 2004. Do comets have chondrules and CAIs? Evidence from the Leonid meteors. *Meteoritics & Planetary Science* 39:1733–1740.

Taylor S. and Brownlee D. E. 1991. Cosmic spherules in the geological record. *Meteoritics* 26:203–211.

Taylor S., Lever J. H., and Harvey R. P. 1998. Accretion rate of cosmic spherules measured at the South Pole. *Nature* 392:899–903.

Taylor S., Lever J. H., and Harvey R. P. 2000. Numbers, types, and compositions of an unbiased collection of cosmic spherules. *Meteoritics & Planetary Science* 55:651–666.

Toppani A., Libourel G., Engrand C., and Maurette M. 2001. Experimental simulation of atmospheric entry of micrometeorites. *Meteoritics & Planetary Science* 36:1377–1396.

Toppani A. and Libourel G. 2003. Factors controlling compositions of cosmic spinerls: Application to atmospheric entry conditions of meteoritic materials. *Geochimica et Cosmochimica Acta* 67:4621–4638.

Whipple F. 1951. The theory of micrometeorites. Part II. In heterothermal atmospheres. *Proceedings of the National Academy of Sciences* 36:687–695.

Yada T. and Kojima H. 2000. The collection of micrometeorites from bare ice of the Yamato Mountains in Antarctica in austral summer of 1998 (abstract #1528). 31st Lunar and Planetary Science Conference. CD-ROM.

Zolensky M. E., Nakamura K., Gounelle M., Mikouchi T., Kasama T., Tachikawa O., and Tonui E. 2002. Mineralogy of Tagish Lake: An ungrouped type 2 carbonaceous chondrite. *Meteoritics & Planetary Science* 37:737–761.