Space on Earth: An Introduction to Meteorites

This material is designed to complement and enhance university-level courses in planetary sciences. The material starts with the birth of the Solar System and moves through the different types of meteorite, their components and what can be learnt from them about the origin and evolution of the Solar System. It concludes with a section about the value of samples returned from space missions and their curation requirements. A brief bibliography of reference materials is included at the end of the document.

This document is complemented by a video presentation available through the EURO-CARES website (http://www.euro-cares.eu/resources).

As well as reading the document, students are encouraged to examine the thin sections of meteorites available through the Virtual Microscope (https://www.virtualmicroscope.org/).

The material has been prepared by Professor Monica Grady of the Open University, Milton Keynes, on behalf of the EURO-CARES consortium. The EURO-CARES project received funding from the European Union's Horizon 2020 research and innovation programme under grant agreement No. 640190

Introduction

Solar System history started some 4567 million years ago with the collapse of an interstellar molecular cloud to a protoplanetary disk (the solar nebula) surrounding a central star, the Sun. Figure 1 is an image of a protoplanetary disk around the newly-forming star HL Tauri. It shows the bright and dense central region where the star is forming, and an increasingly-dispersed disk around the star. Gaps in the disk indicate the potential accretion of a series of planetesimals. This is presumably the type of environment in which the Solar System began.

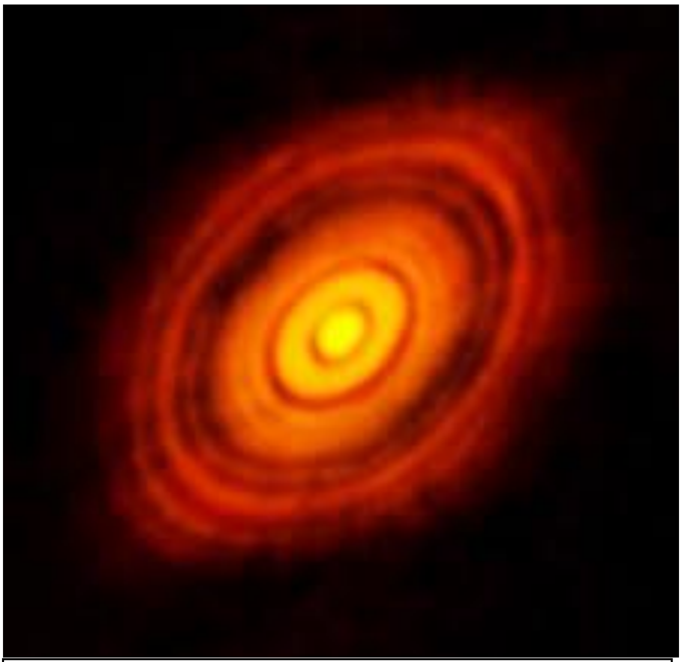

Figure 1: The Protoplanetary Disk of HL Tauri. Image Credit: ALMA (ESO/NAOJ/NRAO), NSF

Evolution of the Solar System continued through a complex process of accretion, coagulation, agglomeration, melting, differentiation and solidification, followed by bombardment, collision, break-up, brecciation and re-formation, then to varying extents by heating, metamorphism, aqueous alteration and impact shock. One of the key goals of planetary science is to understand the primary materials from which the Solar System formed, and how they have been modified as the Solar System evolved. The last two decades has seen a greater understanding of the processes that led to the formation of the Sun and Solar System. Advances have resulted from astronomical observations of star formation regions in molecular clouds, the recognition and observation of protoplanetary disks and planetary systems around other stars, and also from advances in laboratory instrumentation that have led to more precise measurements on specific components within meteorites, e.g., refinement of chronologies based on short-lived radionuclides. Results from meteorites are important because meteorites are the only physical materials available on Earth that give direct access to the dust from which the Solar System formed. Study of meteorites allows a more complete understanding of the processes experienced by the material that resulted in the Earth of today.

Meteorites

Meteorites are pieces of rock and metal, almost all of which are fragments broken from asteroids during collisions. They fall at random over the Earth's surface and have also been identified as components within lunar soils and on Mars' surface. Meteorites are named from their place of find or fall, traditionally after a local geographic feature or centre of population. A complete inventory of the number, type and geographic distribution of meteorites is maintained in the Meteoritical Bulletin (https://www.lpi.usra.edu/meteor/).

Why Classify Meteorites?

Classifying meteorites enables similarities and differences between specimens to be recognised. This, in turn, allows inferences to be drawn about relationships between groups, their origins and the common processes that they have experienced. Over the years, meteorite classification has become a more precise science, partly as a result of the increasing sophistication of the instrumentation available for meteorite analysis, and partly owing to the increasing numbers of meteorites recovered from desert locations. Many schemes for classification have been devised, some with more utility than others, but all schemes, right from the very first descriptions of meteorites, recognised a basic division between stone and iron meteorites. Meteorites can be divided into two main types (Figure 2), according to the processes they have experienced: unmelted (unfractionated, undifferentiated) and melted (fractionated, differentiated).

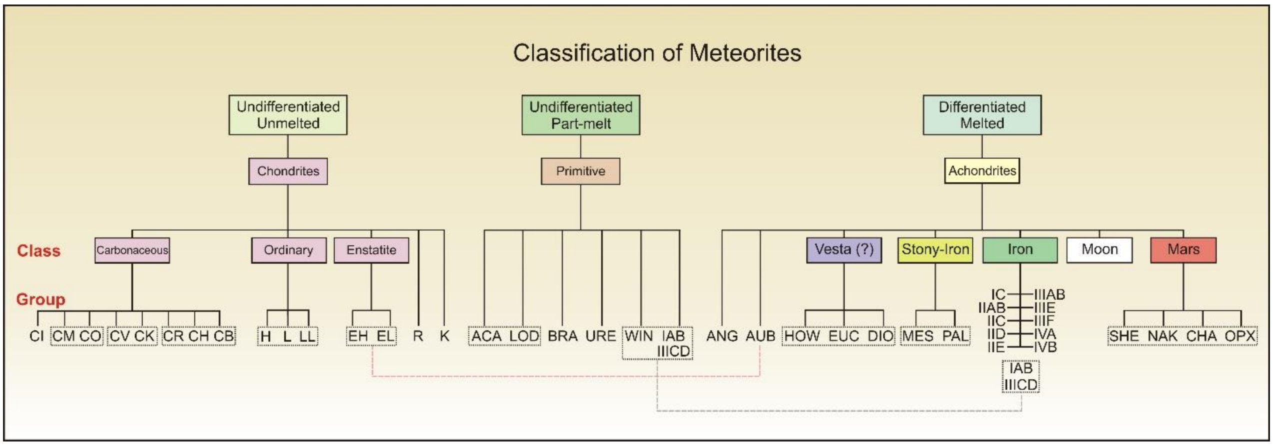

Figure 2: Classification of Meteorite. Image Credit: EURO-CARES/mmg

The unmelted meteorites, or chondrites, are all stones, and in all but the most volatile of elements, have compositions that are close to that of the solar photosphere. Melted meteorites (achondrites) cover a range of compositions from stone, through stony-iron to iron. Bridging between these two major divisions are the primitive achondrites: meteorites that have an unfractionated composition, but textures that indicate they have been strongly heated, if not melted. Both unmelted and melted meteorites are further sub-divided into classes and groups (Figure 2).

Classification of meteorites is one way of identifying materials that might be associated in space and time, e.g., through accretion in closely neighbouring regions of the solar nebula, or having suffered similar processes of heating, melting, differentiation and/or hydrothermal alteration. One of the most widely-accepted methods of classifying meteorites is to determine their oxygen isotopic composition (Figure 3). Whilst this does not identify which specific parent-body a meteorite derives from, it does show which meteorites are related to each other, and which are not. The variations in composition of different meteorite groups are generally taken to be a reflection of primordial nebula heterogeneity, modified in some groups as a result of widespread fluid-solid exchange processes.

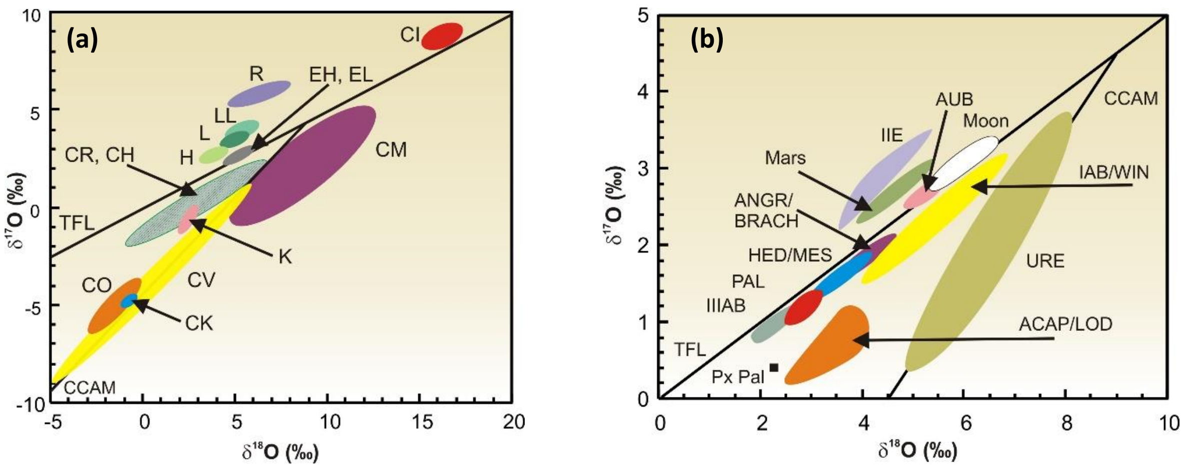

Figure 3: Oxygen isotopic composition of (a) chondrites and (b) achondrites. Data compiled from a variety of sources (see bibliography). Image credit EURO-CARES/mmg.

Despite enormous progress brought about by increasing numbers of meteorites and advances in analytical instrumentation, the classification scheme is incomplete, and there are many meteorites that do not fit comfortably into the framework. There is not always a clear-cut distinction between types: e.g., there are many iron meteorites that contain silicate inclusions related to chondritic meteorites. Clasts and inclusions within meteorites also frequently defy ready assignation to recognised meteorite groups.

Chondrites

We now recognise 15 different chondrite groups (Figure 2). There are eight carbonaceous chondrite groups: CI, CM, CO, CV, CR, CK, CB and CH. The type specimens for the groups are Ivuna, Mighei, Ornans, Vigarano, Renazzo, Karoonda and Bencubbin, respectively. The CH chondrites are not named for a type specimen, but their relatively iron-rich composition is recognised with a label analogous to that of the high iron groups of the ordinary and enstatite chondrites. As well as eight groups of carbonaceous chondrites, there are three groups of ordinary chondrite (H, L and LL), plus the EH and EL enstatite chondrites, and the rumurutiite (R) and Kakangari-type (K) chondrites. The groups are distinguished on the basis of elemental and isotopic chemistry, matrix, metal and chondrule contents and chondrule properties (size, type, etc.). The differences between the groups are primary, i.e. were established as the parent bodies accreted in different regions of the solar nebula. Subsequent to accretion, most meteorites have experienced varying extents of thermal metamorphism or aqueous alteration. These secondary processes occurred on the meteorites' parent bodies, and did not affect the overall composition of the chondrites. Mobilisation and re-distribution of elements such as Fe, Mg and Ca occurred as primary silicates either became homogeneous in composition through heating, or through formation of secondary mineral phases (e.g., clay minerals, carbonates) during aqueous alteration.

Subordinate to classification into the different chemical groups of chondrites is a petrologic sub-classification. This is a numerical (non-linear) scale from 1 to 7 which recognises secondary parent body processes experienced by chondrites, quantifying the extent of dry thermal (petrologic types 3 to 7) or aqueous (petrologic types 1 to 2) alteration experienced after the parent asteroid aggregated. The various meteorite classes have experienced secondary processes to different extents, demonstrated by variety in the range of petrologic types exhibited by the groups. It is also recognised that modification does not cease, even after metamorphism or metasomatism ends: collisions between asteroids lead to brecciation, and the production of shock veins and melt glass. These tertiary effects can be recognised as textural changes within mineral grains. There is a semi-quantitative scale for shock classification based on characterisation of defects in silicate grains (mainly olivine), and ranges from S1 (unshocked) to S6 (highly shocked, to about 75–90 GPa). The final parameter required to describe a chondrite is applied to meteorites that have been found, rather than those observed to fall, as it is a measure of the amount of weathering experienced by the meteorite during its terrestrial lifetime. Weathering rusts metal, and breaks down primary silicates to clay minerals; a semi-quantitative scale to assess the weathering of chondrites runs from W0 (unweathered) to W3 (breakdown of silicates to clays, metal oxidised). The texture of a chondrite results from a combination of two factors: its constituents (i.e., the relative abundances of different components, including matrix, chondrules, metal etc) and the processes that the chondrite (or its parent body) has experienced (thermal metamorphism, fluid alteration).

Components of Chondrites

Chondrites are mechanical mixtures of a variety of components that originated at different times in different regions of the proto-solar nebula. They also contain, to greater or lesser extent, grains that originated prior to the formation of the solar nebula (presolar grains). The relative abundances and sizes of the major chondritic components are essential indicators of the classification of a meteorite.

Calcium- and Aluminium-rich inclusions (CAI)

Calcium- and aluminium-rich inclusions (CAI) are irregularly-shaped objects (Figure 4). As their name implies, they are composed of calcium- and aluminium-rich minerals, such as hibonite, melilite, spinel, etc., as well as fassaite, an aluminium-bearing pyroxene. These minerals are characterized by their refractory nature; their presence implies that CAI formed in high temperature processes (temperatures > 1800 K), probably in a region very close to the newly-formed Sun. The mineralogy and isotopic composition of CAI identified them as objects that formed very early on in the history of the Solar System.

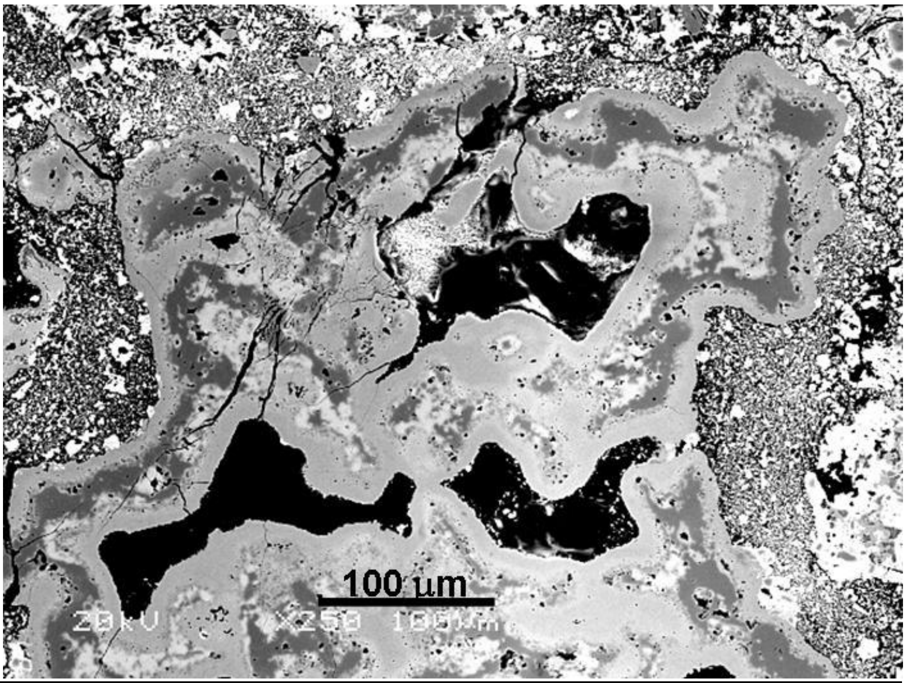

Figure 4: A Scanning Electron Microscope image of a CAI from the Ornans CO3 chondrite

Their irregular shape implies that they have not been through an extensive melting process, and for many years they were thought to have been produced by direct condensation from the cooling nebular gas. Whilst this is presumably correct for some CAI, others have textures that suggest formation from partially molten droplets. CAI are regarded as aggregates of primary nebula condensates that may have experienced several episodes of partial melting and alteration in the nebula prior to accretion into parent bodies; secondary alteration also occurred during subsequent parent body evolution. Age-dating of minerals within CAI has shown that they are the oldest objects in the Solar System, with ages of 4567 Ma. Following their formation (which seems to have occurred on a timescale of < 1 Myr), CAI were transported to the regions in which chondrite parent bodies were aggregating.

CAI are often surrounded by a layered sequence of rims, composed of spinel, perovskite, hibonite, melilite and pyroxene. Each layer is usually composed of only one or two mineral species. The rims follow the outline of a CAI closely, even fractured where the CAI is fractured, so it is assumed that the rims formed in the nebula as part of the CAI, rather than being of parent body origin.

Chondrules

The origin of chondrules is a matter of vigorous and active debate within the meteoritical community. Their spherical to sub-spherical shape and internal texture indicate that they formed in a high temperature process, and that they are quenched droplets of once-molten silicates (Figure 5). Both the location of chondrule formation, and the process that rendered solid grains of interstellar dust molten are uncertain: Scientists have described at least nine different processes that might have been responsible for chondrule formation, including mechanisms such as in the highly energetic wind flowing from the young Sun or in shock waves. It is possible that more than one mechanism was involved in chondrule formation, and that chondrules were produced by different processes at different locations within the nebula at different times.

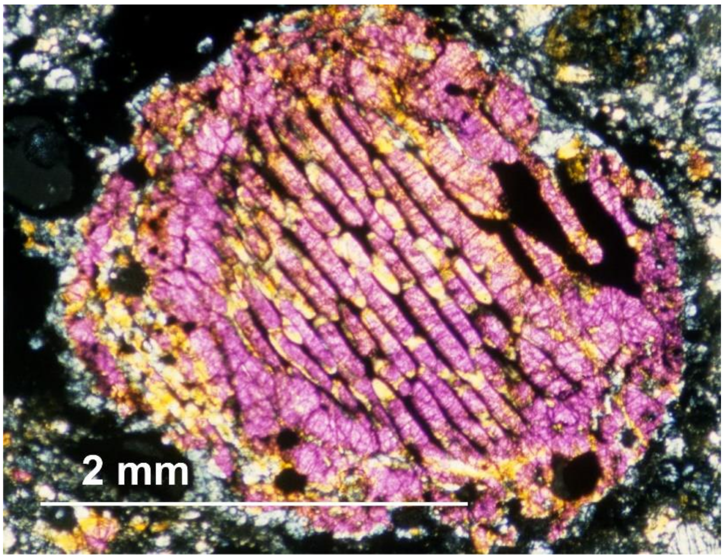

Figure 5: A chondrule from a thin section of the Palymra chondrite pictured in cross-polarised light. Image Credit: NHM/mmg

Chondrules from several different chondritic groups have been dated, giving ages in the range 4562–4567 Gyr. The most ancient age for chondrules has an uncertainty that overlaps with the apparent age of CAI. For many years, it seemed that the formation of CAI and chondrules were separated in time by about 3 Ma. It now appears that CAI and chondrules began forming at almost the same type, but that chondrule formation continued for about 3 million years after CAI formation.

Matrix

The matrix of a chondrite is its groundmass, the material that occurs between chondrules and other discrete objects (including chondrule fragments, CAI and CAI fragments, isolated large silicate, metal and sulphide grains). Matrix is fine-grained (micron to sub-micron) and mainly composed of silicates, although phyllosilicates and organic material are present in some groups of chondrites. Matrix abundance in the different chondrites groups is variable, from CI chondrites that are, in effect, 100 % matrix, to enstatite chondrites with very little matrix at all. As well as forming the groundmass between chondrules, matrix also occurs as rims around chondrules and as individual lumps. The texture of matrix, its degree of recrystallization and the extent to which matrix can be distinguished from chondrules are all parameters employed in assignation of the degree of parent body processing to a chondritic specimen.

Organic Matter

Only 3 of the 15 chondrite groups (the CI, CM and CR chondrites) contain significant quantities of organic matter. It occurs in amounts of 2 to 6 wt. %, and is present as a mixture of aliphatic (both straight- and branched-chain) and aromatic compounds. The main organic constituent is a solvent-insoluble complex network of cross-linked aromatic molecules. At least a part of the organics inventory, characterised by D-, ^{13}C- and ^{15}N-enriched isotopic compositions, was inherited from interstellar space.

Presolar Grains

A volumetrically insignificant, but scientifically critical component within chondrites is their complement of sub-micron to micron-sized interstellar and circumstellar grains. These materials, which include silicon carbide, graphite and diamond, as well as silicates and oxides, are present at levels of a few ppb to ppm and have a variety of origins that pre-date formation of CAI and chondrules. The grains are characterised and classified according to the isotopic composition of their major elements, and the noble gases that are trapped within them. Most studies have been made on presolar grains from CI and CM meteorites, where grain abundance is highest.

Primitive Achondrites

Four groups of meteorites (the acapulcoite-lodranite clan, brachinites, winonaites and ureilites; type specimens Acapulco, Lodran, Brachina, Winona and Novo Urei, respectively) are regarded as primitive achondrites (Figure 2). They have almost chondritic compositions (i.e., they are very little fractionated relative to the sun), but their textures indicate that they have experienced some degree of heating and partial melting. Winonaites are closely related to the silicate inclusions in IAB irons, and may derive from the same parent body, bridging between iron meteorites to chondrites. The characteristics of primitive achondrites may help in the understanding of early differentiation processes in asteroidal bodies. Figure 3 (b) shows the oxygen isotopic compositions of the primitive achondrite groups.

Achondrites

Achondrites are meteorites that formed from melts, and thus have differentiated compositions. Achondrites originally only comprised stony meteorites that had lost a large fraction of their primordial metal content, their name emanating from the observation that such meteorites generally do not contain chondrules. However, the current convention is to regard all differentiated meteorites (stone, iron and stony-iron) as achondrites (Figure 2).

Asteroidal Stone Achondrites

Stony achondrites differ from chondrites in their major element content, especially calcium and similar elements. They have almost no metal or sulphides, and neither do they contain chondrules. They are mainly composed of crystals that appear to have grown from a melt. There are many different groups of achondrites, some of which can be linked together to form associations allied with specific parents. The separate associations have little, if any, genetic relationship to each other. Figure 3 (b) shows the oxygen isotopic compositions of the different achondrite groups.

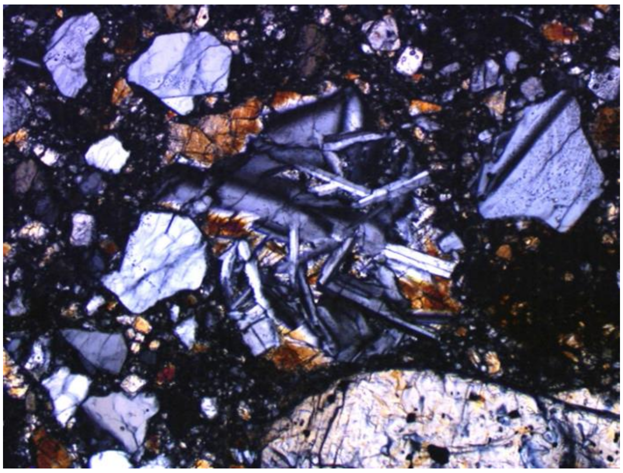

Figure 6: Thin section of the Stannern eucrite. Field of view 4mm. Image Credit NHM/EURO-CARES/mmg

Angrites (type specimen Angra dos Reis) are medium- to coarse-grained basaltic igneous rocks. Although the angrites have similar oxygen isotopic compositions to the HEDs (Figure 3b), they are unrelated. Aubrites (or enstatite achondrites; type specimen Aubres), are highly reduced meteorites with mineralogies and oxygen isotopic compositions similar to those of enstatite chondrites, leading to the suggestion that aubrites might have formed by partial melting of an enstatite chondrite precursor. The howardite-eucrite-diogenite (HED) clan is a suite of generally brecciated igneous rocks ranging from coarse-grained orthopyroxenites (diogenites) to cumulates and fine-grained basalts (eucrites; Figure 6). The howardites are regolith breccias, rich in both solar wind gases and clasts of carbonaceous material. The HEDs all have similar oxygen isotopic compositions (Figure 3b); a strong candidate for the HED parent body is asteroid 4 Vesta.

Stony-irons

The stony-irons are divided into two groups: mesosiderites and pallasites, similar only in their approximately equal proportions of silicate and metal. The two groups have very different origins and histories. Mesosiderites are a much more heterogeneous class of meteorites than the pallasites. They are a mixture of varying amounts of iron-nickel metal with differentiated silicates, the whole assemblage of which seems to have been brecciated.

Pallasites are perhaps the most strikingly beautiful of all meteorites. They are an approximately equal mixture of iron-nickel metal and silicates (predominantly olivine), and the metal forms a continuous network, into which the olivine grains are set.

Iron Meteorites

Iron meteorites are highly differentiated materials, presumed to be products of extensive melting processes on their parents. They are composed of iron metal, generally with between 5–15 wt.% nickel, and account for approx. 5% of all observed meteorites falls. The mineralogy of iron meteorites is dominantly an intergrowth of two phases, the iron-nickel metals kamacite and taenite. Kamacite, or -Fe-Ni, has a body-centred cubic structure and a Ni content < 7 wt.%, whilst taenite, or -Fe-Ni, is face-centred cubic and approx. 15–50 wt.%

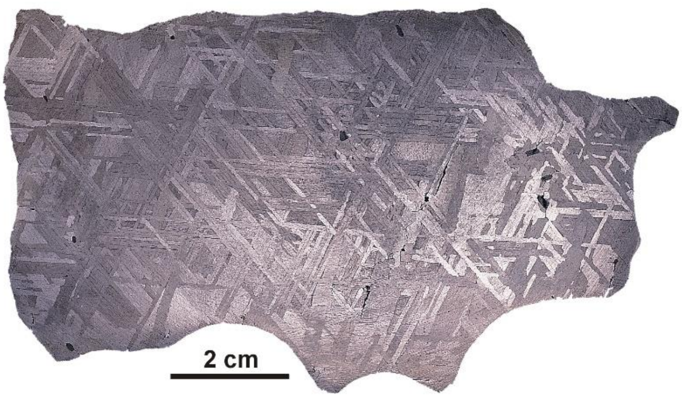

Figure 7: A polished and etched slice of the Henbury iron meteorite. Image credit: NHM

Ni. When polished and etched in dilute mineral acid, irons reveal a distinctive structure, known as a Widmanstätten pattern (Figure 7). This consists of plates of kamacite in an octahedral orientation with interstices between the platelets filled with taenite and plessite (a very fine-grained, sub-micron, intergrowth of kamacite and taenite). The width of the kamacite lamellae is related to the cooling history of the parent bodies. The pattern is named for Aloys J. B. von Widmanstätten, who, at the start of the 19th century, observed the pattern on several iron meteorites, although he probably was not the first to describe the structure.

The coarseness of the Widmanstätten pattern is expressed as the width of the kamacite lamellae (or bandwidth), and iron meteorites were originally classified into 5 structural groups: the coarse, medium and fine octahedrites, ataxites and hexahedrites.

Structural classification of iron meteorites has been succeeded by classification based on metal composition. The irons are sub-divided into 12 different groups (Figure 2) on the basis of nickel and trace element chemistries (Ga, Ge and Ir contents). The relatively tight coupling between Ni and Ga and Ge for ten of the individual groups indicates that the groups represent discrete parents that had completely melted and then solidified by fractional crystallisation. These groups were previously known as the magmatic iron meteorites, although this term is no longer in favour.

Two groups, the IAB and IIICD irons, have a wide range in Ga and Ge abundance with Ni content. They were thought to derive from parent bodies that had not completely melted and were coupled together as non-magmatic iron meteorites that possibly formed during impact processes on their parent asteroids. It is now recognised that silicate inclusions in the IABs are very closely linked to a group of primitive achondrites. If they did not come from the same parent body, they must at least have formed in closely located regions of the solar nebula.

Many irons defy chemical classification, and simply remain ungrouped. On the basis of trace element compositions, it is estimated that iron meteorites might represent samples of at least 70 individual asteroids.

Non-Asteroidal Meteorites

All the meteorites that have been considered so far originated from asteroids. In the early 1980s, it was recognised that there were two groups of meteorites derived from planetary, rather than asteroidal sources.

Lunar meteorites

Lunar meteorites are mostly anorthositic regolith breccias. The first lunar meteorite to be described was ALH A81005 (Figure 8), found in the Allan Hills region of Antarctica. Descriptions of this small (31.4g) specimen were remarkably consistent, and acceptance of its lunar origin was unanimous. The reason that consensus could be achieved so readily was because ALH A81005

Figure 8: The ALH 81005 lunar meteorite. Image credit: NASA

could be compared with the Apollo and Luna samples returned directly from the Moon. The lunar meteorite was identical to Apollo and Luna samples in mineralogy, mineral chemistry and isotopic composition. Cratering of planetary surfaces by asteroidal impact had been considered as an important process for modifying planetary surfaces, but from the dynamics of such a process, the ejection of large amounts of material was thought to be unfavourable. Identification of ALH A81005 as lunar showed that material could indeed be removed from the Moon and land on Earth.

Martian meteorites

There have been many missions to Mars since the first successful fly-by of Mariner 4 in 1965. Imagery and other data from orbiting satellites and landed craft have increased our knowledge of Mars and our understanding of its formation, the relative chronology of the major events that shaped its surface, and the mineralogy of the surface materials. A valuable source of information about Mars, complementary with data returned from space missions, is the class of meteorites that were ejected from Mars' surface by impact.

Figure 9: The EET 79001 martian meteorite. Image credit: NASA

The existence of lunar meteorites was readily accepted because they could be compared directly with material returned by the Apollo and Luna missions. Because astronauts have not yet visited Mars, we have no samples returned directly from our neighbouring planet to provide a benchmark against which potential martian meteorites could be compared. However, once the idea that meteorites could be removed from the Moon by impact was accepted, then the idea of meteorites from Mars gained ground. The breakthrough in realisation that there were rocks from mars in the world's meteorite collections came following analysis of another meteorite collected in Antarctica, EET 79001 (Figure 9).

The tale of how a group of meteorites was accepted as martian is too long and complicated to be included here (but see bibliography). A summary of the evidence for a martian origin is as follows: the meteorites are younger than asteroidal meteorites (the youngest is ~ 165 Myr old) so they cannot be from the asteroid belt, but must be from a volcanically-active body (i.e., a planet). Oxygen isotope composition shows that all the meteorites come from the same parent body (but do not indicate which body that is). Finally, some of the meteorites contain gas trapped within pockets of melt glass. The gas has the characteristic composition, in terms of CO_2, N_2 and noble gases, as Mars' atmosphere (as determined by orbiting spacecraft). It also has the same $^{40}Ar/^{36}Ar$ isotopic composition as Mars' atmosphere, which is very different from that of Earth's atmosphere. The gas was concluded to be a sample of Mars' atmosphere trapped when the meteorite was excavated from Mars' surface by impact. Hence the meteorites are from Mars.

There were originally only three sub-groups of martian meteorites, the type specimens of which were named Shergotty, Nakhla and Chassigny. On this basis, the martian meteorites used to be referred to as 'SNCs', but usage of this term is now discouraged since, as indicated above, several more sub-groups have been recognised.

Age-dating of Meteorites

The main events that led to the formation and evolution of the Solar System (CAI and chondrule formation, aggregation, crustal melting, differentiation and core formation) can be followed by several different radiometric age-dating chronometers. An underlying assumption of the dating technique is that the parent isotope was distributed homogeneously through the meteorite (asteroid) formation regions.

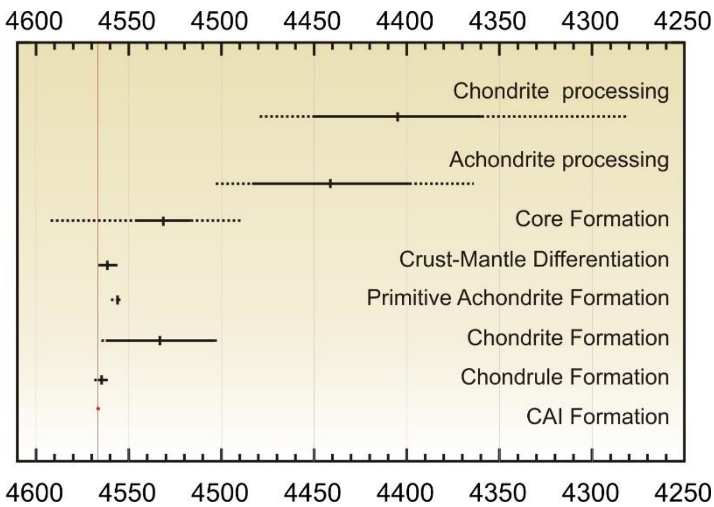

Figure 10: Solar System chronology. Data compiled from a variety of sources (see bibliography). Image credit: EURO-CARES/mmg

Several isotope systems with different half-lives are used to measure different events within Solar System history. The absolute age of a sample is obtained using the U-Pb system. The two separate isotopes of uranium (^{235}U and ^{238}U) decay at different rates to two different isotopes of lead (^{207}Pb and ^{206}Pb, respectively). The ratio of the two daughter isotopes, $^{207}Pb/^{206}Pb$, can be measured very precisely, and if several components from a single sample are measured, the resulting isochron yields an age. The Pb-Pb age obtained in this way is an absolute age. Short-lived radioisotope systems provide ages that are relative to the Pb-Pb absolute age; if sample that has a Pb-Pb age is also dated by a short-lived chronometer, then the two ages can be compared directly and the timescales inter-calibrated. Absolute ages have now been determined for many meteorite classes and meteoritic components.

Formation Ages

The oldest components in meteorites are the CAI, with ages (determined by U-Pb dating) of 4567 ± 0.6 Myr. This age can be taken as a baseline from which the date of formation of other meteoritic components can be measured. The decay of the short-lived radionuclide ^{26}Al has been applied extensively as a fine scale chronometer for determination of the relative ages of early Solar System events. Because Al and Mg are both lithophile elements, they are insensitive to metal-silicate fractionation processes, and so the ^{26}Al-^{26}Mg chronometer is used to constrain the formation timescale of primitive rocky materials. This has led to the establishment of a relative timescale for the production of CAI and chondrules. The 'canonical' model for accretion rationalised that the short-lived radionuclides were produced externally to the presolar nebula (probably in a supernova), injected into the collapsing dust cloud and incorporated into the CAI, etc, on timescales shorter than that of the radionuclide half-life. For example, the presence of ^{26}Mg within CAI shows that the CAI formed whilst ^{26}Al was still "live" in the solar nebula, i.e., agglomeration took place over a very short timescale, < 1 Myr. The ^{41}Ca-^{41}K chronometer implies even more rapid formation of CAI, with an interval between nucleosynthesis and agglomeration of < 0.3 Myr. The most abundant component within chondrites, chondrules, shows little evidence for live ^{26}Al; measurements on aluminium-rich chondrules using the Al-Mg chronometer imply that chondrules formed after CAI production, and that chondrule-formation took place over an extended period of time, commencing very shortly after CAI formation, and completing within < 5 Myr. This is compatible with the Pb-Pb ages of chondrules, which suggest that the onset of chondrule formation was almost simultaneous with CAI formation, but which then continued for several million years after CAI formation was completed.

Planetary Melting and Differentiation

Following planetesimal accretion, silicate melting occurred, the heat source for which was presumably radioactive isotope decay combined with impact-induced heating. Subsequent crystallization results in partitioning of elements between different mineralogical components; a suitable radioisotope chronometer for measuring the age of crystallization of a melt (i.e., the time when elemental exchange could no longer occur between phases) is one where parent and daughter isotopes have different cosmochemical behaviours, such that, for example, the parent partitions into one silicate mineral (e.g., olivine), whilst the stable isotope of the daughter element moves into a second mineral, e.g., pyroxene. One such radiometric chronometer is the ^{53}Mn-^{53}Cr system ($t_{1/2}$ = 3.74 Myr), which has been extensively employed to date the timing of crystallization of early-formed igneous rocks, such as angrites. Concordance between the Mn-Cr and Pb-Pb isotope systems in angrites has been used to pin the two chronometers together, allowing absolute age differences to be inferred on the basis of measurements of short-lived radioisotope decay products.

The final stage in the gross evolution of a planetary embryo is melting of large volumes of material, leading to metal-silicate differentiation, segregation of metal and core formation. Again, these processes can be followed by application of a suitable chronometer, for example the hafnium-tungsten system (^{182}Hf-^{182}W; $t_{1/2}$ ~9 Myr), which has been used to date metal-silicate differentiation on the Earth, Moon, Mars, and asteroidal parents. Both hafnium and tungsten are highly refractory elements that are present in approximately chondritic relative abundances in bulk planets and planetesimals. However, during metal-silicate differentiation (or core formation), the lithophile Hf segregates into the silicate fraction while the siderophile W segregates into the metallic core. If this differentiation event occurs within ~5 half-lives of ^{182}Hf, ~45 Myr, then its timing can be determined by measuring the excess of the daughter product, ^{182}W. This has enabled new estimates for the timing of metal-silicate differentiation on the terrestrial planets and planetesimals. Iron meteorites started to form around the same time as CAI formation. The onset of core formation is now thought to have occurred within the first ~3 Ma and ~13 Ma from the beginning of Solar System formation for Vesta and Mars respectively, with those of the Earth and Moon forming some time later.

Thermal and Shock Histories

The thermal and shock histories of meteorites are illustrated by changes in mineralogy and texture. The time when such changes occurred can be determined from measurement of the noble gas inventory of a specimen, because these species are mobile, and readily lost during secondary processing. As long as a system remains undisturbed (i.e., not shocked or metamorphosed), the abundance of noble gases that have accumulated in a specimen is a record of how much time has elapsed since its final disturbance. This gas retention age might be dating the crystallization of a rock from a melt, or the era when thermal metamorphism occurred, or the timing of a shock event such as catastrophic break-up of a parent object. The noble gas species most significant for determination of thermal or shock history is ^{40}Ar, yielding a ^{40}K-^{40}Ar age, which is the age of closure to argon, either since crystallization (in an undisturbed sample), or since the last thermal or shock event in a disturbed sample. An equivalent age is obtained by a variant on the technique, whereby neutron irradiation of a specimen converts ^{39}K into ^{39}Ar, giving an ^{40}Ar-^{39}Ar age. Measurement of ^{40}Ar-^{39}Ar ages of lunar samples and meteorites have shown clusters in ages that are related to impact and break-up events.

Collecting and Curating Meteorites and the value of Sample Return Missions

There are almost 60, 000 meteorites in the worlds' collections of meteorites; only about 1300 of them were seen to fall – meaning that the overwhelming majority of meteorites have been contaminated to greater or lesser extent during their terrestrial residence. This does not imply that such meteorites are not extremely important to study, it just means that there may be an extra level of complication associated with interpretation of results from any investigation. Hence the value of sample return missions, and the necessity for appropriate receiving and curation facilities once samples come back to Earth.

Currently, we have samples that have been collected directly from three Solar System bodies: the Moon, a comet and an asteroid, as well as material trapped from the Solar Wind. Lunar samples (almost 300 kilograms in total) from the Apollo and Luna missions have been available for analysis for over 30 years. It was comparison with Apollo samples that confirmed the validity of a lunar source for the first lunar meteorite to be identified.

Over the next decade, there are clear opportunities for Europe to lead a sample return mission to the Moon, and to collaborate with other space agencies on sample return missions to asteroids and to Mars and its moons (Phobos and Deimos). ESA, as well as national and other international space agencies, have several missions under study to these bodies. It is essential that a sample receiving and curation facility is considered as a critical element of the mission architecture and that its planning and design requirements are fully incorporated during the earliest phases of planning for each sample return mission. Previous work has indicated that from site selection to full-readiness for receiving Mars samples takes 8 - 11 years.

Planetary Protection

One of the most important issues surrounding sample return missions is the requirement for Planetary Protection (PP). This guides the design of a mission, aiming to prevent biological contamination of both the target celestial body and, in the case of sample-return missions, the Earth. The Committee on Space Research (COSPAR) has the mandate from the United Nations to maintain and promulgate the planetary protection policy. Planetary protection is essential to preserve our ability to study the astrobiologically-interesting planets and moons of our Solar System by preventing contamination with terrestrial micro-organism or organics and thus removing the possibility of false-positive results (forward PP). The second aspect of planetary protection aims to protect the Earth's biosphere from extra-terrestrial agents, which might be harmful if released into the Earth environment (backward PP). Both aspects have been considered, forward PP on samples collected and then returned, and backward PP during transport and curation phases.

COSPAR defines five planetary protection categories with subcategories dependent on the target of the mission and the type of mission (fly-by, orbiter or lander). All missions which will return extra-terrestrial samples to Earth for further analysis belong to category V. Depending on the origin of the extra-terrestrial material a category V mission can be an unrestricted Earth return mission (e.g. samples from the Moon) or restricted Earth return mission (e.g. samples from Mars or Europa).

Once returned to Earth, samples have to be stored under specific conditions (depending of their origin) so they remain as pristine as possible. At the same time, for restricted missions, the Earth environment must also be protected from potential hazards. Currently, worldwide, no single facility exists that allows containment of restricted materials, as would be required for a sample receiving facility for materials returned from objects such as Mars. Since it is impossible to foresee the actual risk factor of returned samples, the facilities need to have the most stringent containment level presently afforded to the most hazardous biological entities known on Earth. The infrastructure, procedures, protocols and instrumentation, sample handling, as well as staff training shall all be adapted to PP requirements.

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