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The Near Earth Objects: possible impactors of the Earth/Les astéroïdes geocroiseurs : impacteurs potentiels de la Terre

Meteorites: samples of NEOs in the laboratory

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Abstract

Meteorites can be considered as samples of near-Earth objects (NEOs), and as such they may be useful for inferring the properties of the latter, including those that may encounter the Earth in the future. This article reviews the main properties of meteorites which can be of interest in NEO research. We first briefly recall the characteristics of a meteorite fall and the biases it introduces in the passage from the near Earth meteoroid population to the meteorite population. We then describe in more detail the mineralogical and chemical composition of the various classes of meteorites. The relations between meteorites and asteroids that can be inferred from reflectance spectroscopy, and the porosity of meteorites are treated in Sections 4 and 5, respectively. The last section deals with meteorite ages, with emphasis on the cosmic ray exposure age. ***To cite this article: C. Perron, B. Zanda, C. R. Physique 6 (2005).***

Résumé

Météorites : échantillons de NEOs au laboratoire. Les météorites peuvent être considérées comme des échantillons d'objets proches de la Terre ('NEOs') et en tant que telles elles peuvent être utiles pour prévoir les propriétés de ceux-ci, notamment de ceux qui pourraient tomber sur Terre dans le futur. Cet article passe en revue les principales propriétés des météorites qui peuvent présenter un intérêt dans la recherche sur les NEOs. On rappelle d'abord les caractéristiques de la chute d'une météorite et les biais qu'elle introduit dans le passage de la population des météoroïdes proches de la Terre à la population des météorites. On décrit ensuite plus en détail la composition minéralogique et chimique des différentes classes de météorites. Les Sections 4 et 5 traitent respectivement des relations entre météorites et astéroïdes telles qu'on peut les déduire de la spectroscopie en réflectance, et de la porosité des météorites. Les ages des météorites, et plus particulièrement l'age d'exposition au rayonnement cosmique, sont abordés dans la dernière section. ***Pour citer cet article : C. Perron, B. Zanda, C. R. Physique 6 (2005).***

Keywords: NEO; Meteorites; Reflectance spectroscopy

Mots-clés : NEO ; Météorites ; Spectroscopie en réflectance

1. Introduction

When wondering about NEOs, especially about the possible chemical and physical properties of their materials, for scientific reasons or for the purpose of evaluating and mitigating the risks they present to human civilization, it soon comes to mind that

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we already have samples of them in the laboratory. Indeed, natural objects of extraterrestrial origin fallen onto the surface of the Earth, or, as the official I.A.U. definition has it, “*meteoroids which have reached the surface of the Earth without being completely vaporized*” [1], meteorites were small NEOs until their encounter with the Earth terminated their interplanetary voyage. As such, there is little doubt that they can be of some help in characterizing both the NEO population as a whole and individual NEOs, in particular those which may pose a threat to the Earth, but to what extent?

In what follows, we review the main characteristics of meteorites, emphasizing those relevant to NEO research. Topics with no direct bearing on this subject will not be treated here in any detail, whatever their scientific interest (physico-chemical conditions and chronology of the early solar system, organic matter, presolar interstellar grains, etc.).

2. Meteorite falls

As an incoming ‘small’ meteoroid (i.e. smaller than say 10 m in size) reaches the top of the atmosphere, friction against the air decelerates it and the conversion of kinetic energy to heat can raise the temperature of the projectile surface and of the surrounding air to several thousand degrees Kelvin. The characteristics of the fall depend on a number of factors, including the velocity, mass and shape of the meteoroid, the orientation of its trajectory, and the density and strength of its material. The lower the mass, the more efficient the air drag. Small particles, in the millimeter size range, are stopped in the upper atmosphere and the corresponding meteors do not last more than a fraction of a second; these are the familiar ‘shooting stars’. Larger masses penetrate deeper into the atmosphere and make a longer and more spectacular show, in particular in the case of meteorite-producing events. The initial velocity in this case is generally about 15–20 km/s. A fireball appears at a height of about 100 km. The outermost part of the bolide matter is melted, vaporized and ejected, leaving a trail of vapor, dust and ionized atoms. The mass of the meteoroid thus decreases continuously, a phenomenon known as ablation. In addition, because of the very strong tensions to which it is submitted the bolide very often breaks into several, many, or sometimes thousands of fragments (Fig. 1). At a height of about 20 km, the velocity has been so much reduced – it is now about 3 km/s – that incandescence can no longer be sustained, ablation ceases and the fireball switches off. In general, no more than a few seconds have elapsed since it appeared, but in exceptional cases (nearly horizontal trajectories) it has been seen for up to about 40 s [2]. The trajectory now bends over to nearly vertical, and the meteorite finally reaches the ground with its velocity of free fall (one or a few hundred m/s). If fragmentation occurred in flight, the larger fragments being less efficiently braked travel farther, so that on the ground, fragments are found distributed in an elliptical strewnfield, with the largest pieces at one end and the smallest ones at the other.

Because of this fiery arrival, meteorites have a typical aspect: their surface is black, smooth, with rounded angles (Fig. 2). However, as the fall is so short, and the heat generated is in a large part carried away by the vaporized material and the surrounding air, the interior of the meteorite fortunately suffers very little: the black vitrified layer that surrounds it, called the

Fig. 1. Photograph of the Peekskill meteor (9 October 1992), testifying to the fragmentation of meteorites in the atmosphere. More than 70 fragments can be counted, of which only one has been retrieved, after it passed right through a car trunk. (photo courtesy of S. Eichmiller/Altoona Mirror).

Fig. 2. Four specimens of the L'Aigle meteorite, fallen on 26 April 1803, from the MNHN collection. The largest one has a mass of 6 kg. One clearly sees, where it is broken, that the heat-affected zone, the black fusion crust, is only a very thin layer (a few tenths of a mm thick).

fusion crust, is not usually thicker than a fraction of a millimeter. A few mm away from the fusion crust, the meteoritic materials are essentially unaffected by the fall. Meteorites are thus real samples of NEOs. However, ablation and fragmentation in the atmosphere introduce some discrimination depending on the strength of the meteoroid material. In the atmosphere, a meteoroid preferentially breaks along weak points such as fractures and voids. Highly fractured meteoroids are thus more efficiently fragmented and destroyed during atmospheric flight. In addition, different kinds of meteorites withstand the atmospheric braking differently: out of the 45 known meteorites having a total mass larger than one ton, only 6 are stony meteorites, while, as we shall see below, they represent some 93% of all falls; similarly, out of the 31 meteorites having at least one fragment weighing more than one ton, only two are stony [3]. This demonstrates that stony meteorites in general suffer much more from ablation and fragmentation than iron meteorites. It is thus certain that the meteorite population does not faithfully reproduce that of NEOs. Fragile, friable meteorites, as for instance the CI chondrites, must be underrepresented in our collections, while, conversely, the stronger ones, as iron meteorites, must be overrepresented. In this perspective, the fact that we do not have cometary meteorites appears a likely result of the selection by the atmosphere – unless we do have some, but have been unable to recognize them up to now [4,5].

Bodies larger than about 10 m are little decelerated in the atmosphere, which leads to catastrophic outcomes. For this reason, there are no intact meteorites larger than a few meters. Stony meteoroids in the size range 10–100 m explode in the atmosphere [6–10]; the higher the mass and the mechanical strength, the lower the height of the explosion. Such events occur nearly 10 times per year, without being noticed from the ground [11]. This is probably what happened on 30 June 1908 above the region of the river Tunguska, in Siberia, but in this case the explosion took place low enough to destroy approximately 2000 km^2 of forest. Still larger stony meteorites, and iron meteorites over ~ 10 m reach the ground with almost their cosmic velocity. Most of their kinetic energy transforms into heat and they are vaporized upon impact, together with a much larger quantity of target rocks, leading to the formation of an impact crater [12].

In some cases, analyses of crater samples revealed a contamination by meteoritic matter and identified the nature of the projectile. Small craters (less than about 1 km in diameter) are made by iron meteorites [13], as stony meteorites of the corresponding size are destroyed in the atmosphere. In this case, fragments of iron meteorite are frequently found scattered around the crater. For larger craters, the nature of the projectile seems to reflect more closely the distribution of classes observed for small meteorites, and no meteoritic fragments survive [13]. Some 170 impact craters are known on Earth [14], and many more on all the solid bodies from Mercury to Neptune's moon Triton, making impact cratering a major geological process in the solar system. One of the largest known crater on Earth, the Chicxulub crater in Mexico with a diameter of ~ 200 km, results from

Non-differentiated meteorites

Class	FF	Number	Group
Carbonaceous chondrites	4.1%	6	**CI**
		14	**CM**
		5	**CO**
		6	**CV**
		2	**CK**
		3	**CR**
		0	**CH**
		1	**Bencubbinites**
		3	**C-ungr**
	0.1%	1	**R**
	0.1%	1	**K**
Ordinary chondrites	79.8%	76	**LL**
		8	**L/LL**
		365	**L**
		2	**H/L**
		332	**H**
Enstatite chondrites	1.7%	8	**EL**
		9	**EH**

Differentiated meteorites

Group	Number	FF	Class
IAB, IC, IIAB, IIC, IID, IIE, IIF, IIIAB, IIICD, IIIE, IIIF, IVA, IVB, others	48	4.9%	*Irons*
Pallasites	5	0.5%	
Mesosiderites	7	0.7%	
Eucrites	30	5.8%	*Basaltic achondrites*
Diogenites	11		
Howardites	16		
Ureilites	5	0.5%	
Angrites	1	0.1%	
Aubrites	9	0.9%	*Enstatite achondrites*
Acapulcoites Lodranites	2	0.2%	*Primitive achondrites*
Brachinites	0	<0.1%	
Winonaites	1	0.1%	
Anorthosites Basalts	0	<0.1%	*Lunar meteorites*
Shergottites	2	0.4%	*Martian meteorites*
Nakhlites	1		
Chassignites	1		
Others	0		

Fig. 3. The main classes and groups of meteorites. The major distinction is between differentiated and non-differentiated objects. A few intermediate groups however exist (primitive achondrites) and there are mineralogical, chemical and isotopic similarities between enstatite chondrites and enstatite achondrites, which indicate a genetic relationship. The number of *observed falls* in each group is indicated as it is more representative of the nature and frequency of meteorites falling onto the Earth than the much larger total number of meteorites (which is biased by a number of factors). The fall frequency (FF) for each class is also indicated. The total number of falls taken into account in this figure is 981. Data from [3,76]. Note that a few objects (4 observed falls and about 60 more finds) do not come from asteroids but from the Moon and from Mars.

the impact of a meteorite about 10 km in size, 65 Ma ago [15,16]. There is no doubt that such large impacts cause ecological catastrophes on the scale of the whole planet. Sixty five Ma ago is the time of the Cretaceous-Tertiary boundary crisis, when a large fraction of the living species, including the dinosaurs, disappeared.

3. Mineralogy, chemistry and classification

The traditional classification sorted meteorites into stony, iron and stony-iron meteorites on a purely descriptive basis. A better understanding of the origins of meteorites has now allowed us to classify them into a number of classes and groups which reflect these origins (Fig. 3). The most important distinction is between differentiated and non-differentiated meteorites (chondrites). Differentiated meteorites come from asteroids and larger planets which, like the Earth, experienced interior melting mostly due to the decay of freshly synthesized radioactive nuclei integrated to the solar nebula and subsequently to the planets. Differentiation separated an initially homogeneous body (with a composition reflecting that of the nebula) into several layers with different compositions: a dense metallic liquid (containing Fe, Ni and all the siderophile elements) sank towards the centre of the planet to generate a core, whereas a less dense silicate melt (containing Na, Al and Ca) erupted at the surface, forming a basaltic crust. The residual material, essentially ferro-magnesian, formed the mantle. In differentiated bodies, no single sample is representative of the global composition of the planet, whereas chondrites still reflect the composition of their parent asteroids, inherited from the solar nebula. Fig. 4a shows a comparison between the chemical abundances measured in chondrites of

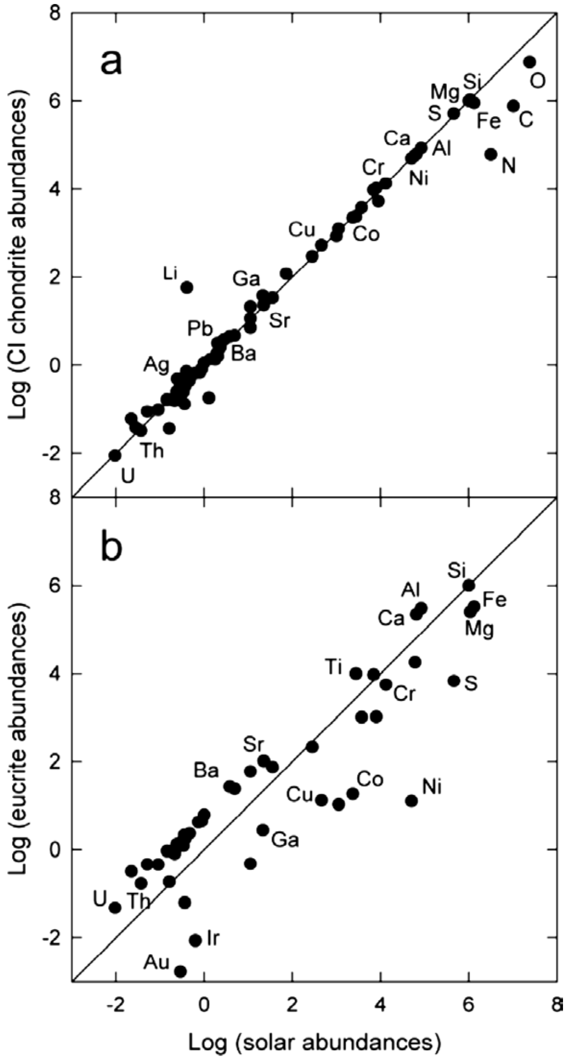

Fig. 4. Abundance of the elements in meteorites, as a function of their abundance in the solar photosphere. (a) CI chondrites; (b) eucrites. Note the logarithmic scales. CI data from [17], eucrite data from [77] and the review by [78].

the CI group and in the solar photosphere. Because of convection in the Sun, the solar photosphere is supposed to be representative of the whole Sun, which, in turn, comprises >99% of the mass of the solar system. The agreement between solar and chondritic abundances is excellent over 8 orders of magnitude, except for H, He and other rare gases (not plotted), C, N, and O, all volatile elements which have been only partly incorporated into solid bodies, and for Li, which is destroyed by nuclear reactions in the Sun. As CI chondrite abundances are measured with a greater precision than those of the solar photosphere and for a larger number of elements, they form the basis for the standard solar system abundances [17]. Fig. 4b, on the other hand, shows the equivalent diagram in which abundances in differentiated meteorites (asteroidal basalts) are compared to abundances in the Sun. The basalt is enriched in Ca, Al and the lithophile elements (i.e. silicate-forming elements), whereas it is depleted in all the siderophile elements which have sunk into the complementary core. Rocks from differentiated planets or asteroids thus lost any memory of the original chemical composition of their parent body through melting, and also of its structure, which would have reflected the nebular environment in which they formed. These rocks are however precious for studying planetary processes, in particular those which took place in small planets in the early solar system.

The meteorite groups listed in Fig. 3 were originally established on mineralogical and chemical criteria, as discussed in more detail below. A newer and extremely powerful criterion is now available: each meteorite group has a characteristic oxygen isotopic signature [18]. As shown in Fig. 5, these signatures cannot be explained by mass-dependent fractionation alone: a mixing from at least 2 different oxygen reservoirs, containing different amounts of ^{16}O (as they lie on a slope 1 line), is required as

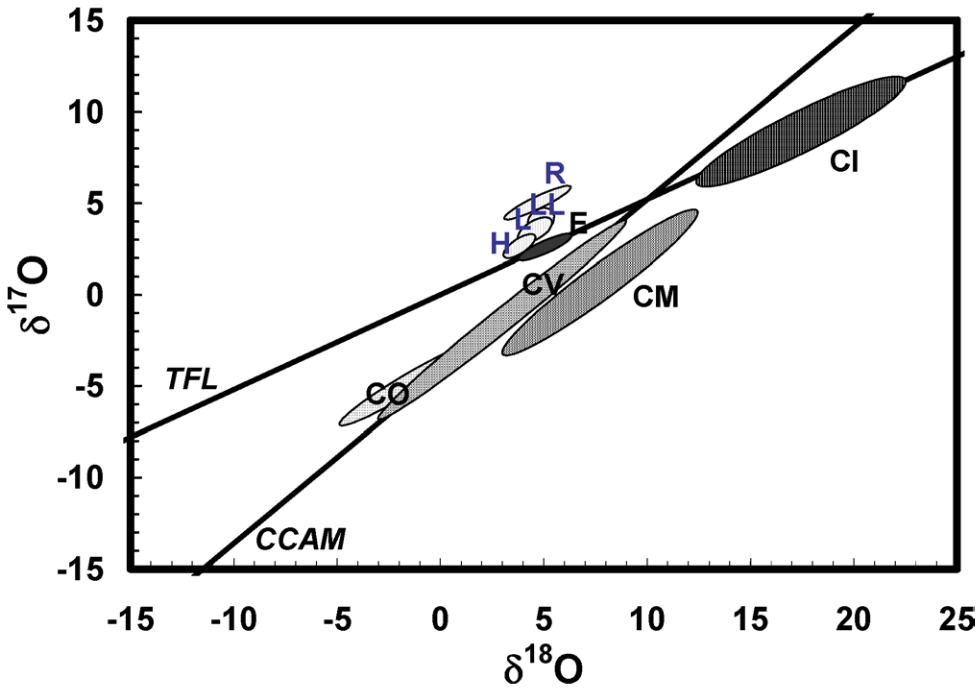

Fig. 5. Oxygen isotope signatures of major chondrite groups. For convenience, the small variations of oxygen isotopic compositions are described in terms of their deviation in parts per thousand from a terrestrial standard, SMOW (Standard Mean Ocean Water):

$$\delta^{17}O = \{[(^{17}O/^{16}O)_{sample}/(^{17}O/^{16}O)_{SMOW}] - 1\} \times 10^3,$$
$$\delta^{18}O = \{[(^{18}O/^{16}O)_{sample}/(^{18}O/^{16}O)_{SMOW}] - 1\} \times 10^3.$$

Each chemical group of meteorites has its own oxygen signature (with some amount of overlap between groups). For simplicity, only the major chondrite groups are displayed here and EH and EL chondrites are grouped as 'E'.

Two types of isotopic fractionations are involved: mass-dependent and mass-independent. Mass-dependent fractionations (usually the result of temperature, coordination or kinetic effects) move the samples along a slope $\frac{1}{2}$ line. Except for some very special cases, all terrestrial samples are thus restricted to the $\frac{1}{2}$ slope line labeled 'TFL' (terrestrial fractionation line), and each chondrite group is characterized by its $\Delta^{17}O$ (the excess of $\delta^{17}O$ relative to the terrestrial fractionation line). In meteorites, mass-independent effects are also present, the origin of which is still not properly understood. Most individual components from CV chondrites lie along a slope ~ 1 line known as 'CCAM' (carbonaceous chondrite anhydrous minerals) which was suggested to result from the addition of pure ^{16}O to these samples [79]. Ordinary chondrite chondrules, on the other hand, all fall in the OC domain, i.e. above the TFL. After review by [18].

well. Although the nature and significance of these distinct oxygen reservoirs are still debated, oxygen isotopes nevertheless provide an extremely useful tool to classify doubtful samples, reveal new groups, and point to possible genetic relationships between mineralogically or chemically distinct meteorite groups [18].

3.1. Non-differentiated meteorites

Non-differentiated meteorites are called *chondrites* because an extremely striking feature of most of the primitive ones is that up to 80% of their crystals are arranged in more or less spherical structures, with diameters in the range 50 μm to several mm, called *chondrules* (see Fig. 6). The space between chondrules is filled by a matrix made of very fine grains (< 1 μm) of similar minerals, where unmodified. As discussed below, the origin of chondrules is still an open question. It is, however, clear that they were made from the crystallization of individual droplets and that they carry information relative to the environment of the forming Sun and planets. Chondrites are made of silicates, iron-nickel alloys (Fe–Ni) and iron sulfide (troilite, FeS). The most common silicates are olivine ($[Fe,Mg]_2SiO_4$), pyroxenes ($[Fe,Mg,Ca]SiO_3$), and feldspar (solid solution of $[Na,K]AlSi_3O_8$ and $CaAl_2Si_2O_8$). Many other minerals have been identified in chondrites, usually only in minor amount except for phyllosilicates (clay-like minerals), which are dominant in the chondrites that are considered the most primitive in terms of their chemical compositions.

Although all chondrites have a composition close to that of the Sun, only one group, the CI chondrites, actually matches the composition of the solar photosphere, as displayed in Fig. 4a. All other chondrite groups are chemically fractionated with

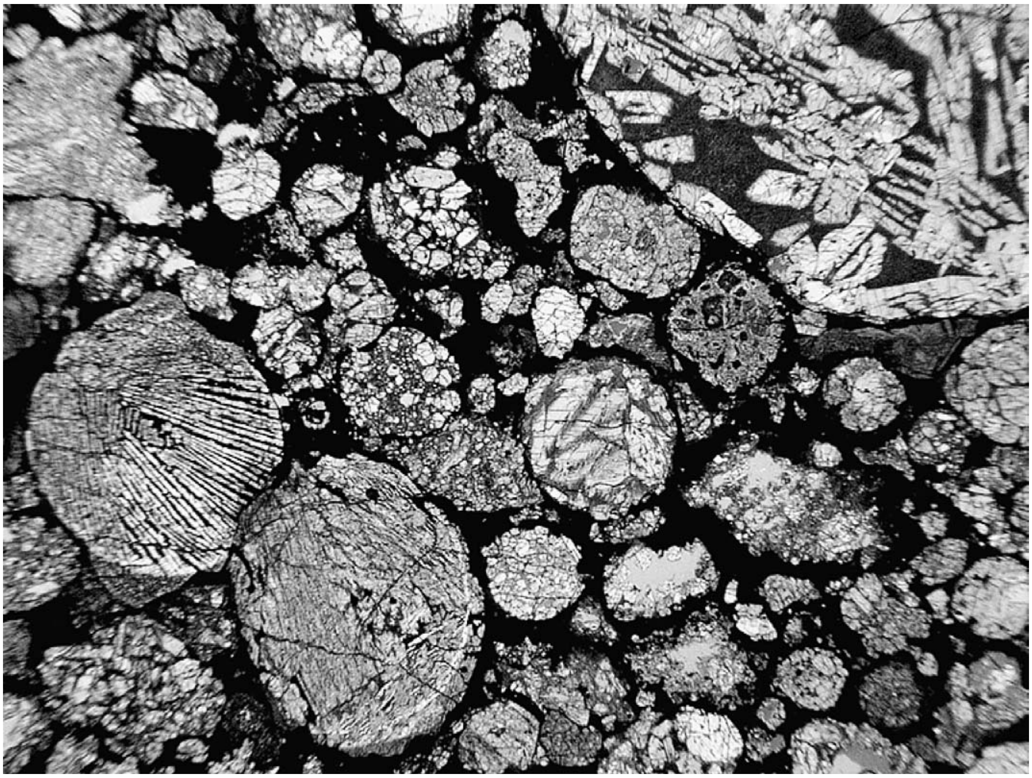

Fig. 6. Thin section of the Semarkona chondrite (LL3.0), showing abundant chondrules and chondrule fragments of various textures and sizes (note the large chondrule only partially visible in the upper right corner). The opaque material filling space between chondrules is the fine-grained matrix. Transmitted light. Width of photo: 7 mm (photo courtesy of M. Bourot-Denise).

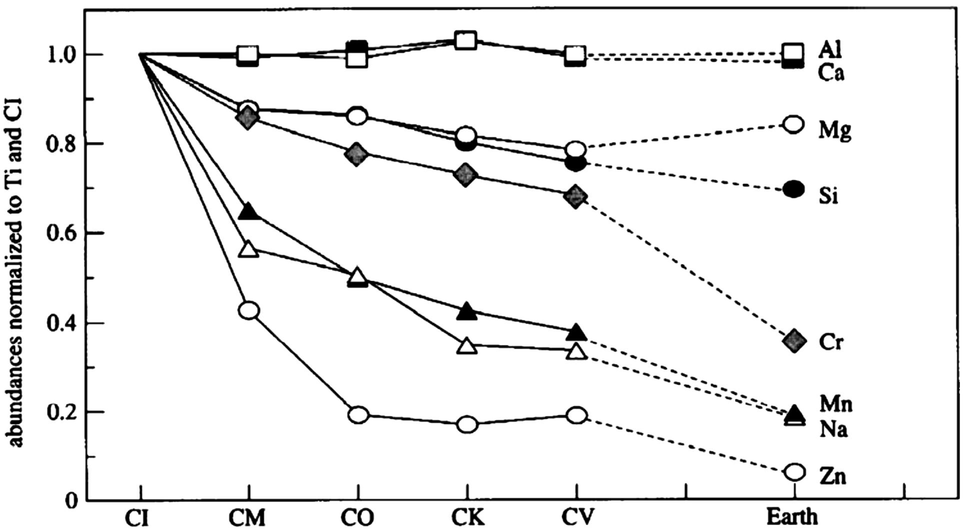

Fig. 7. Abundance of a few elements in various carbonaceous chondrite groups compared to the solar composition of CI chondrites and normalized to Ti. All chondrite groups (and the Earth's mantle) exhibit chemical fractionations compared to CIs and the Sun. The more volatile an element is (Zn, Na), the more fractionated it can be, compared to the refractory elements Ti, Al and Ca. This suggests the fractionation resulted from a high temperature process involving evaporation and/or condensation [80].

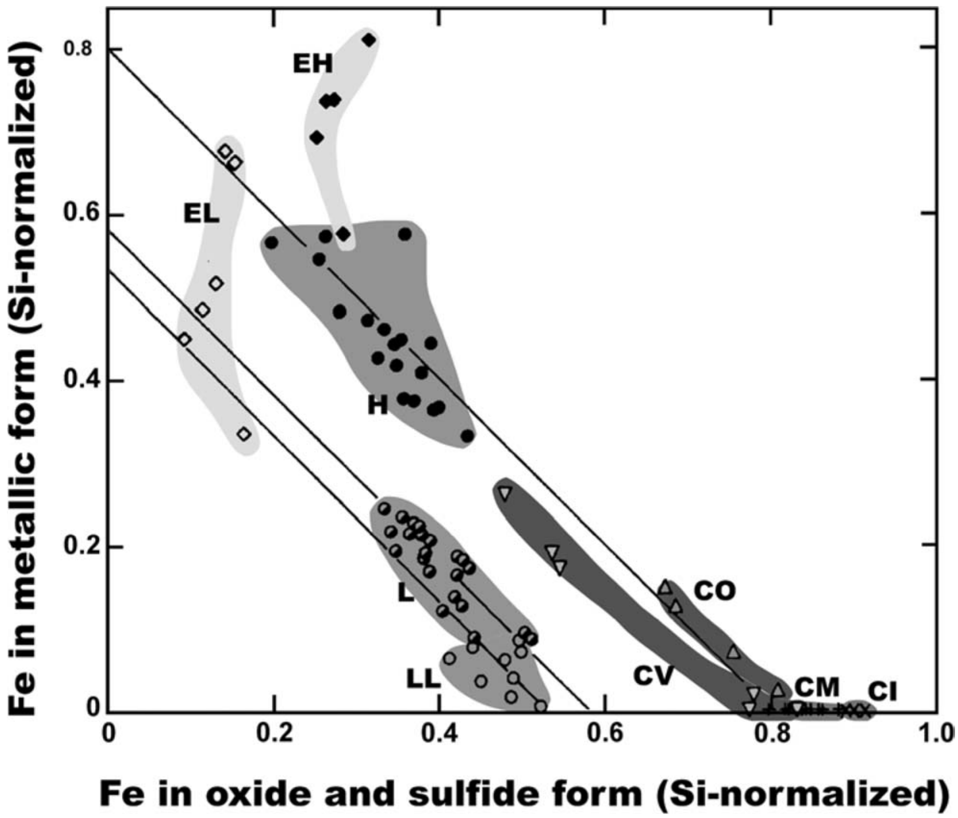

Fig. 8. Chondrite groups were originally defined based on their total Fe content and its oxidation state. In this modified version of the Urey and Craig diagram (after [81]), total Fe is constant along the diagonals, increasing from lower left to upper right. Although CIs have no metal, only EHs have more total Fe. As for their other chemical properties and the abundance of their chondrules, CMs are intermediate between CIs and the other carbonaceous chondrite groups, having less total Fe and a small amount of metal. EH chondrites are enriched in Fe and in volatiles in general with respect to CIs.

respect to CIs (carbonaceous chondrite compositions are shown in Fig. 7). The variability of their Fe content and of its oxidation state has been one of the criteria for defining these groups (Fig. 8) which we now know to correspond to distinct reservoirs based on their oxygen isotopic signatures [18]. Three main classes of chondrites were recognized early on: the ordinary (OCs, most common), enstatite (ECs, so reduced that some lithophile elements such as Mn or Ca are found in sulfides) and carbonaceous chondrites (CCs), some of which contain organic matter up to a few %. In Fig. 3, these chondrite classes and the groups in which they were sub-divided are arranged in roughly decreasing order of degree of oxidation from top to bottom. Carbonaceous chondrites are the most oxidized: many do not contain any Fe–Ni metal, all Fe being in oxidized form, in silicates or other minerals (e.g. magnetite). Enstatite chondrites are the most reduced: they contain large amounts of Fe–Ni metal, but almost no Fe in their silicate which is nearly exclusively enstatite, a purely magnesian pyroxene ($MgSiO_3$). Ordinary chondrites are intermediate, with Fe–Ni alloy content increasing from LL to L to H. These differences probably reflect different local conditions in the solar nebula at formation time.

Although chondrites have a primitive chemical composition, they have nevertheless experienced various amounts of planetary processing. Even if the heat from the decay of radioactive species was insufficient to melt their parent asteroids after accretion, it induced some internal transformations due to the diffusion of atoms between minerals that were not in chemical equilibrium. Asteroidal materials were also subjected to important shocks as part of their accretion, which meant brief but intense releases of energy. Finally, water must have circulated in some of the asteroids, altering their minerals. To account for all these transformations, chondrites have been sorted into 6 petrologic types [19] (Fig. 9). Types 3 to 6 correspond to increasing effects of thermal metamorphism. Type 3s are the least affected: their minerals have variable compositions, in particular the $Fe/(Fe + Mg)$ ratio varies among silicates, they are said to be unequilibrated chondrites; chondrules are abundant and sharply defined. Type 6s are the most metamorphosed: their minerals have uniform compositions, they are equilibrated chondrites; the chondrule texture has been almost entirely erased by recrystallisation, and chondrules can hardly be recognized in these chondrites. Types 1 and 2 correspond to negligible planetary thermal processing, but considerable effects of alteration by water on

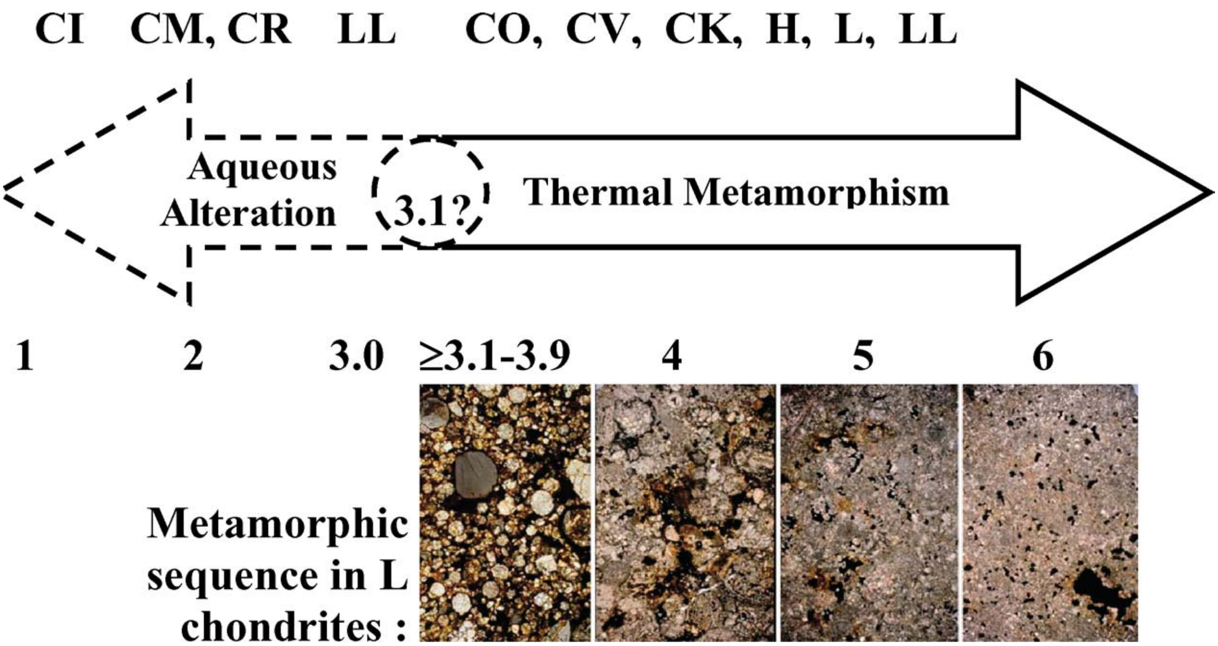

Fig. 9. Chondrites experienced various levels of secondary processing on their parent asteroids. The petrographic types describe both aqueous alteration and thermal metamorphism as the two processes appear to be negatively correlated, at least in the first order. Unfortunately no totally pristine material seems to exist: the lower type 3 chondrites were both reheated and altered to a moderate level. The main groups found in each type are listed at the top. Type 1 meteorites (CI) consist almost exclusively of alteration minerals, also abundant in type 2s (CM + CR). From type 3s to type 6s (all other groups), reheating-induced diffusion has gradually driven mineral compositions and textures towards equilibrium, erasing the original limits of chondrules and matrix. The chondrules remain well delimited throughout type 3s and they can still be distinguished in type 4s. In the type 5 sample shown here the phantom of a chondrule can still be made out, but no chondrules virtually remain in a type 6. (All transmitted light views are at the same scale, field: 4.7×7 mm.)

the parent asteroids. Type 1 chondrites, the most primitive from a chemical point of view, are almost entirely composed of very fine-grained hydrous minerals, in particular phyllosilicates analogous to clay minerals found on Earth, which are the products of aqueous alteration of anhydrous silicates. Chondrules are totally absent from these chondrites, but it is not known whether they never existed there, or were destroyed by the action of water. Not all petrologic types are represented in the various chemical groups. For instance, there are no types 1 and 2 among the ordinary and enstatite chondrites, CIs (which contain 17% water [20]) are mostly type 1, CMs (which contain 10% water [20]) are mostly type 2, CVs and COs are type 3.

Chondrule textures imply that they formed as small, independent melted droplets (Fig. 6). Simulation experiments closely reproduce these textures and suggest formation temperatures in the range 1400–1600 °C, and heating times of minutes to hours [21]. However, the physical phenomenon responsible for the (transient) energy inputs necessary to make chondrules is still unknown, although many have been suggested. Most of the hypotheses refer to processes taking place in the solar nebula, but planetary processes, a priori less probable, are periodically revived [22]. This ignorance is all the more irritating in that the high abundance of chondrites (85% of all meteorite falls) and the high proportion of chondrules in chondritic matter (up to 80%, in ordinary chondrites) strongly suggest – but do not prove – that chondrule forming was a common process and one of those which most affected the solid matter of the early solar system. Let us mention two of the most recently developed models of chondrule formation: the nebular shock model [23,24] and the x-wind model [25,26]. In the former, solid chondrule precursors are melted first by radiation from an approaching shock wave, then by friction with the gas accelerated to supersonic speeds by this shock. The model accounts for the thermal history of chondrules as inferred from observations and experiments as well as for some other properties of chondrules. The origin of the shocks remains an open question, although the proponents of the model suggest gravitational instabilities in the disk. In the x-wind model, chondrules are melted by impulsive protosolar flares at the inner edge of the accretion disk, close to the protoSun, and launched to planetary distances by an x-wind outflow. This model is in a large part based on observations of young stellar objects. It offers explanations not only for the formation of the chondrules themselves but for more general characteristics of the chondrites. It has not been shown, however, that it can match the chondrule thermal history in detail or the complementarity of composition that seems to exist between chondrules and matrix in a given meteorite.

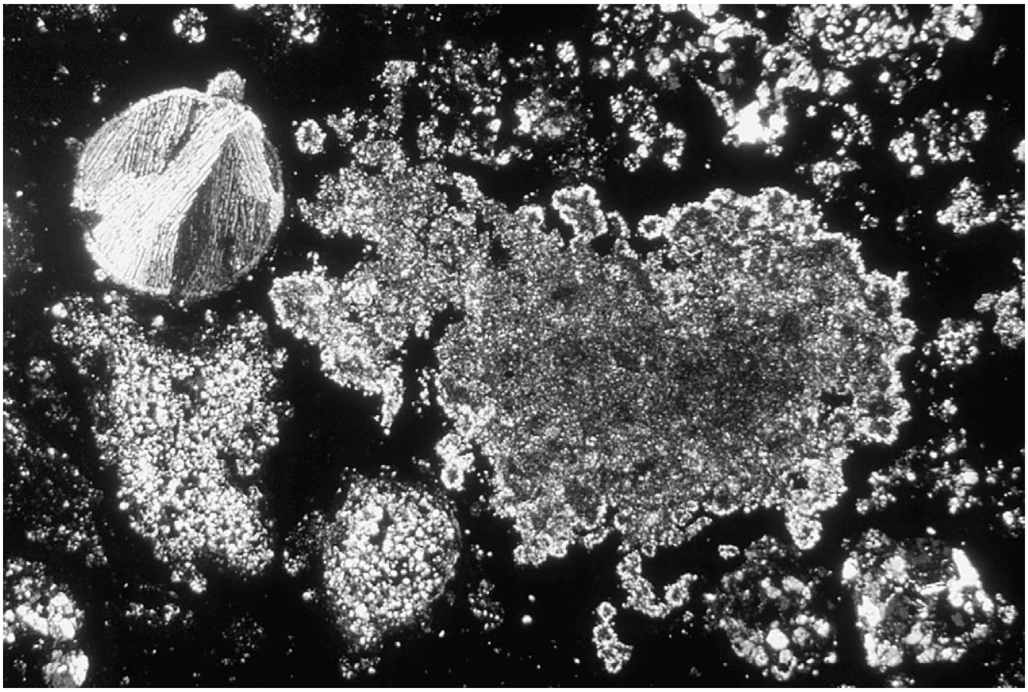

Fig. 10. A polished thin section from the CV3 chondrite Allende is seen in transmitted polarized-analyzed light. A chondrule is visible at the upper left and a refractory inclusion occupies most of the picture. Width of photo: 2 mm (photo courtesy of M. Bourot-Denise).

Refractory inclusions are another important type of object formed at high temperatures (Fig. 10). They are found in most carbonaceous chondrites, especially in CV3s where they are the largest (few mm to cm), and, rarely, in ordinary and enstatite chondrites. They are assemblages of refractory, Ca- and Al-bearing minerals and are designated for that reason calcium-aluminum-rich inclusions (CAIs). CAIs appear to be the first solids to have formed in the solar system. They keep memories of very ancient events, in the form of isotopic anomalies for many elements, which result from nucleosynthetic processes that occurred before the birth of the Sun, or just after it within the solar nebula. They were made by condensation and/or evaporation in the very first instants of the solar system, and subsequent melting in many cases. In the x-wind model, CAIs are formed by a mechanism similar to chondrules, but closer to the Sun, and they are heated for longer times, including during transport in the x-wind. As a consequence, many elements evaporate, leaving a refractory residue rich in calcium and aluminum oxides and silicates.

Carbon is present in small amounts in all chondrites, and represents up to about 4.5% by mass of the CI and CM chondrites [20], the most carbon-rich of all. In these chondrites C is mainly in organic matter, the most abundant form of which is a poorly characterized macromolecular component, but many molecules are also present, such as aliphatic and aromatic hydrocarbons, carboxylic and amino acids, etc. [27]. Isotopic analyses of hydrogen in this organic matter show that it is strongly enriched in its heavy isotope, deuterium (^{2}H or D) [28], which is difficult to explain by processes that have taken place in the solar system. This suggests that at least part of this organic matter is of interstellar origin, i.e. was synthesized in the interstellar medium before the birth of the Sun, was thereafter incorporated into materials of the solar nebula, and survived up to now in some meteorites. It is generally accepted that ion-molecule reactions that occur at low temperatures in interstellar molecular clouds can yield considerable hydrogen isotope fractionation, and indeed D-rich molecules have been detected in the interstellar medium [29].

3.2. Differentiated meteorites

Unlike chondrites, differentiated meteorites have compositions widely different from that of the Sun (Fig. 4b). They result from total or partial melting of their parent asteroids, followed by a sequence of fractional crystallization. The extreme case is iron meteorites, which are made of only two major elements: iron and nickel. They consist of Fe–Ni alloys, with variable amounts of inclusions of iron sulfide, graphite, silicates and other minerals. On the basis of the concentrations of Ni and trace elements (especially Ga, Ge and Ir) [30], they are sorted into 13 groups (Fig. 3) corresponding to different parent asteroids, but many ($\sim$13%) do not fit in any of these groups. Most irons are believed to be fragments of metallic cores resulting from extensive early melting of asteroids.

Pallasites are made of Fe–Ni and cm-sized crystals of olivine in about equal proportions. They probably originate from the interface between the metallic core and the rocky mantle of differentiated asteroids. Howardites, eucrites and diogenites (HEDs) are collectively known as basaltic achondrites, because they resemble a basalt (lava) crystallization sequence. Eucrites are made of pyroxene and feldspar, diogenites of pyroxene, and howardites of agglomerated fragments of eucrites and diogenites. They are magmatic rocks ejected from the crust of a differentiated asteroid. As discussed below, there is evidence that this asteroid is 4 Vesta. Mesosiderites are irregular mixtures of Fe–Ni metal and of silicates similar to those found in eucrites and diogenites. They are thought to result from a collision involving the metallic core of an asteroid and its silicate surface [31] or that of another one [32]. Angrites are a very small group of magmatic meteorites which crystallized very early in the solar system history and are most probably from a parent body different from that of the eucrites. Ureilites are distinguished from all other achondrites by their high carbon content (1.5–3%). Their origin is far from clear, various hypotheses include differentiation of a carbonaceous asteroid and/or an asteroid collision. Aubrites consist almost entirely of enstatite. They have an igneous texture, but are almost surely related to the enstatite chondrites as indicated by their mineralogy and their oxygen isotopes (Fig. 5). The rare Acapulcoites-Lodranites and Brachinites are often referred to as primitive achondrites, although they are quite different from each other (the former are Fe–Ni metal-rich, the latter almost entirely made of olivine). In spite of a highly recrystallized texture, they have a nearly chondritic composition, except for losses of elements like S and Na which disappeared in small melt fractions.

4. Relation with asteroid classes

Reflectance spectroscopy, which provides information on the mineralogical composition of asteroid surfaces, also allows a direct comparison between meteorites and asteroids. Similarities have been noted for a long time between the spectra of some asteroid classes and laboratory spectra obtained on various powdered meteorites, suggesting possible genetic links between meteorite and asteroid classes. However, progress in this area has been rather slow, and very few unambiguous associations have been established between given asteroid classes and meteorite groups. For instance, the proposed associations of one of the seven carbonaceous chondrite groups with one asteroid or one asteroid class are still speculative [33–37]. Probably the most successful suggestion in this domain has been the association of the basaltic achondrites with 4 Vesta [38,39] and the small Vestoids [40]. Although there are opponents of this theory [41,42], the similarity of the reflectance spectra of HEDs with those of Vesta, together with the existence of the Vestoids and of a huge crater on Vesta [43], have made a strong case for Vesta being the 4th identified body of the solar system (with the Earth, the Moon and Mars) from which we have samples.

A longstanding issue has been that the ordinary chondrites, which make up 80% of all meteorite falls on Earth, have no or very few [44] spectral equivalents in the asteroid belt. One of the suggested explanations of this puzzle has been that some space weathering process makes asteroid surfaces darker and redder and absorption bands weaker. This hypothesis has now gained support from both laboratory experiments and asteroid observations. Nanophase reduced iron has been observed on the surface of lunar grains, that accounts for the modifications of the optical properties of lunar soils [45]. These tiny metal blebs were plausibly produced through bombardment of the surface of the Moon by micrometeorites and energetic particles. Formation of such iron nanoparticles, accompanied by reddening of reflectance spectra, has also been observed on olivines irradiated by nanosecond pulsed laser, simulating micrometeorite impacts [46]. Fresh craters on asteroids 951 Gaspra and 243 Ida are blueish compared to the rest of the asteroid surface [47]. A positive correlation has been found between the color and the age of S asteroid families [48]. As a whole, these observations make it very likely that the parent bodies of the ordinary chondrites are among the S-complex asteroids, and that the spectral differences are, at least in part, due to a time-dependent transformation of the asteroid surfaces.

The situation is quite comparable when one considers NEOs instead of main belt asteroids. Exact matches between NEO and ordinary chondrite spectra are rare [49–51], but the visible and near-infrared wavelength spectra of a number of NEOs span the range between those of ordinary chondrites and those of S asteroids [52,53], giving more support to S asteroids being the ordinary chondrite parent bodies, with surfaces more or less modified by space weathering. Spectral and chemical measurements of S-asteroid 433 Eros by the NEAR-Shoemaker spacecraft are somewhat ambiguous, but the most likely analogs of Eros are ordinary chondrites [54]. Thus, asteroids made of materials similar to ordinary chondrites may be abundant among NEOs. Some ambiguity remains in the association of carbonaceous chondrites with NEOs: K-types are linked either to CV3 [53] or CO3 chondrites [51], some C-types better fit with CV3 than with CI1 or CM2 chondrites [51], contrary to what is generally accepted.

5. Physical properties: density and porosity

As far as impact risks are concerned, porosity is probably one of the most important properties of NEOs to be determined, along with the mass of the objects, as it will in a large part control their behavior during mitigation actions or atmospheric flight. Meteorites offer the opportunity to determine the porosity of various kinds of NEOs, with the high precision accessible

Table 1
Bulk density (ρ_B g/cm^3) and porosity (%) of a selection of meteorites and asteroids

Meteorites					Asteroids		
Name	Group	ρ_B	Porosity	Ref.	Name	Class	ρ_B
Orgueil	CI1	1.58	35	[56]			
		2.06	13*	[59]			
Murray	CM2	2.08	29	[58]	1 Ceres	G	2.12
Murchison	CM2	2.37	16	[58]	2 Pallas	B	2.71
Allende	CV3	2.92	24	[56]	253 Mathilde	C	1.3
		2.85**	18**	[58]			
Kainsaz	CO3	2.98	12	[58]			
Ordinary chondrites		2.6–3.7	5–30	[56–58]	243 Ida	S	2.6
					433 Eros	S	2.67
Juvinas	Euc.	3.03	<1	[56]			
Millbillillie	Euc.	2.86	11	[58]	4 Vesta	V	3.44
Tatahouine	Dio.	3.09	7	[58]			
		3.26	0	[58]			
Brenham	Pal.	4.72	5	[56]			
Augustinovka	IIIB	6.99	2	[56]	16 Psyche	M	2.0

* Calculated using Orgueil grain density from [55].
** Average of measurements on 5 different samples.
Euc.: eucrite; Dio.: diogenite; Pal.: pallasite.
Asteroid data are from the review by Britt et al. [60].

in the laboratory. Although porosity determination is not one of the routine measurements commonly made on meteorites, a few such studies have been published (see review [55] and references therein). The porosity of a sample is usually determined from the difference between its bulk volume and its grain volume. The bulk volume – the total volume, including pore spaces – is generally measured by immersing the sample in a powder made of tiny glass or plastic beads. The grain volume – the volume of the mineral grains, excluding pore spaces – is obtained by immersing the sample in helium. A measurement of the mass of the sample allows us to also calculate the bulk density and the grain density. There are however some reservations. As already mentioned, we only have 'small' meteorites: the largest known is only about 3 m in its longest dimension, and the largest stony meteorite has a diameter only slightly larger than 1 m. Moreover, porosity measurements have been made on much smaller samples still, with a mass less than 2 kg, often no more than a few grams or tens of grams. Cracks and voids of a size comparable to or larger than the size of these samples are of course not taken into account in these measurements. Even smaller cracks may be underrepresented in the studied samples, as meteoroids are likely to fragment preferentially along these during atmospheric flight. The measured porosity of meteorites should thus be considered 'microporosity', a lower limit of the porosity of asteroids, which will in general include a 'macroporosity' component due to large-scale voids, and one should expect to find higher porosity values for asteroids than for meteorites. The difference between the density of an asteroid and that of meteorites made of the same material will provide a measure of its macroporosity and thus key information on its internal structure.

Table 1 presents a (somewhat arbitrary) selection of bulk density and porosity values of individual meteorite samples. Many more data can be found in the original references [56–59] and in the review by Britt and Consolmagno [55]. Carbonaceous chondrites can have rather high porosity, but measurements on different samples of the same meteorite (Orgueil in Table 1) or different meteorites of the same group (Murray and Murchison in Table 1) can yield quite different values. This latter remark is also true for the basaltic achondrites. This may result from porosity heterogeneity or may indicate experimental problems. Few meteorites of these classes have been measured and measurements on more samples and more meteorites are clearly needed. The porosity distribution of ordinary chondrites, which are by far the most studied meteorite class for density and porosity, strongly peaks around 10%, but has a tail extending up to ∼30% [57]. Not surprisingly, irons and stony irons have very low or zero porosity.

Also included in Table 1 is a selection of bulk density values of asteroids, mostly from the main belt [60]. A comparison of bulk densities of meteorites and asteroids of possibly related classes shows that the largest asteroids have a very low macroporosity, some others have a significant macroporosity (∼20%), but are not much more porous than the most porous meteorites, and a third group, with a macroporosity in excess of 30%, probably consists of loosely bound rubble piles [60]. This kind of information is about totally missing for NEOs, and it would be highly desirable to extend density determinations to them, and, at the same time, to make more measurements of meteorite densities and establish clearer links between asteroid classes and meteorite groups.

6. Meteorite ages

Different ages can be defined for meteorites, depending on the method used for the determination. The most relevant to the NEO topic is the cosmic ray exposure age.

6.1. Cosmic ray exposure age: duration of interplanetary journey

When a meteoroid is extracted from its parent body by an impact, it starts being exposed to irradiation by cosmic rays, from which it was protected before by overlying rocks. Energetic cosmic ray particles induce nuclear reactions in the meteoroid on a depth scale of the order of 1 m, breaking atomic nuclei and creating new ones, called cosmogenic nuclides. This irradiation stops when the meteorite reaches the surface of the Earth, whose atmosphere protects it against cosmic rays. A measurement of the amount of cosmogenic nuclides in a meteorite sample yields the duration of the exposure, i.e., the time it took the meteorite to travel from its parent body to the Earth. This presentation is somewhat oversimplified, as it appears that a relatively large fraction of the meteorites were close enough to the surface of their parent body to be already irradiated before their ejection, and thus have a more complex irradiation history. In some cases, the duration of the two exposures can be determined, but meteorites with simple irradiation history are clearly more convenient for statistical studies of exposure ages. Useful cosmogenic nuclides are radioactive nuclides and isotopes of rare gases as only these are initially present in amounts low enough for the small addition due to cosmic ray-induced nuclear reactions to be detected.

Exposure ages of ordinary chondrites are mostly in the range 2–80 Ma, those of carbonaceous chondrites tend to be shorter, usually less than 10 Ma, while those of irons are mostly comprised between 200 and 1000 Ma and extend to more than 2 Ga [61]. In a study of the exposure age distribution of ordinary chondrites, Marti and Graf [62] find, in addition to the early recognized peak at 7 Ma in the H-group distribution, another peak at 33 Ma and peaks at 28 and 40 Ma in the L-group and at 15 Ma in the LL-group distributions. They conclude that these histograms reflect individual impact events and are not consistent with a continuous delivery of asteroidal fragments to the Earth. Similarly, the exposure age distribution of the HED achondrites shows evidence for two impact events at 22 and 39 Ma ago [63].

All these ages are different from the much shorter lifetimes, of order 1 or a few Ma, calculated for transfer of objects through resonances from the asteroid belt to the Earth [64]. This puzzle has been solved by taking into account the Yarkovsky effect [65,66]. The latter is a force acting on small spinning bodies in the solar system, due to anisotropy in the absorption and emission of thermal radiation [66]. Under the action of the Yarkovsky effect, asteroid fragments can slowly drift from the site where they have been ejected from their parent asteroid to a resonance, which then quickly sends them to Earth-crossing orbits. In this picture, exposure ages of meteorites essentially reflect the time spent drifting in the asteroid belt, not the duration of the transfer from the asteroid belt to the Earth. The Yarkovsky effect is weaker on bodies with higher thermal conductivity, which accounts for the longer exposure ages of irons than for stones. However, not all features of exposure age distributions can be explained this way [67].

It should be remarked that exposure ages are not directly comparable to calculated lifetimes. The latter correspond to the durations of the journey of all fragments from one impact event, while exposure ages correspond to the durations of the journey of the fragments from all possible impact events that arrive on Earth at one and the same moment, i.e. now. The fact that we receive now, for instance, H chondrites from an event which took place 33 Ma ago does not mean that many more objects from this event were not received in the past, and the near absence of LL chondrites with exposure ages less that 7 Ma may just reflect the absence of major LL chondrite-producing impacts in the last few Ma. Fossil L chondrites have recently been found in 480 Ma old sediments [68,69]. Their number indicates a meteorite flux at Earth about two orders of magnitude higher than the present flux. Their exposure ages are within the range 0.1–1 Ma [70]. Admittedly, these meteorites come from an exceptionally large impact, whose traces are seen in a large fraction of the L chondrites in the form of an important degassing dated at about 500 Ma ago [71]. Fragments from this event may have been ejected with high velocities and directly injected into a resonance, hence their very short exposure ages. Nevertheless, it may be interesting to look for fossil meteorites with terrestrial ages corresponding to the various peaks in the exposure age distributions.

6.2. Crystallization age: age of the solar system

The clocks used to determine the age of a rock are pairs of nuclides present in the rock, one of which is produced by the radioactive decay of the other, for instance ^{87}Rb–^{87}Sr, ^{40}K–^{40}Ar, ^{238}U–^{206}Pb, etc. The transformation of the parent-nuclide (P) into the daughter nuclide (D) follows a law whose time dependence is well established:

$$N_D(t) = N_P(t)(e^{\lambda t} - 1),$$

where $N_D(t)$ is the number of atoms of D produced at time t, $N_P(t)$ is the number of remaining atoms of P, and λ is the decay constant of P. By measuring N_D and N_P, one can thus determine the time t, called the age of the rock. This is complicated by the fact that one has to take into account the quantity of D initially present at $t = 0$, but there are technical solutions to this problem. Now what does $t = 0$ means, in other words, when did the clock start? The clock started at the closure of the system, i.e. at the time when no more addition or removal of either P or D occurred in the sample studied, except for what is due to radioactive decay. This happened when the temperature of the sample became low enough to prevent atomic diffusion. The (crystallization) age of a rock is thus the time elapsed since its temperature passed below the closure temperature of the system. The crystallization ages of asteroidal meteorites, whether differentiated or not, are all close to 4.5 Ga [72].

The CAIs mentioned above are present in primitive meteorites, whose temperature always remained low. Their age is thus the time elapsed since the end of the high temperature event in which they formed. Using the U–Pb clock, an age of 4.567 Ga has been determined, with an uncertainty of only 0.7 Ma [73]. No older material has been found, and this age of 4.567 Ga is taken as the age of the solar system.

6.3. Terrestrial age

After arrival on Earth, meteorites are shielded from cosmic rays by our atmosphere. Radioactive cosmogenic nuclides thus decay within a meteorite without being produced any more. A comparison of their concentration in a collected meteorite with that normally found in a fresh fall gives the time elapsed since arrival on Earth, called the terrestrial age. Meteorites collected in Antarctica have terrestrial ages up to 2 Ma for stones and 5 Ma for irons [74,75].

7. Conclusion

The large diversity of the meteorites we have in our collections, which has been impressively increasing during the last decade, together with the possibility of making all sorts of measurements on them with sophisticated instruments in the laboratory, are a real chance for those interested in inferring the properties of near-Earth objects. However, deducing NEOs' properties from the properties of meteorites is not straightforward, owing to a number of biases due in a large part to the interaction of the incoming meteoroids with the atmosphere and the surface of our planet. Meteorites are small, since big ones are destroyed upon arrival; large scale features, in particular large voids and fractures are thus not seen in our NEO samples. The traversal of the atmosphere at high speed makes a selection on the type of material that reaches the ground; some types of meteorites are thus underrepresented in our collections with respect to the NEO population, or even totally missing, as seems to be the case for cometary meteorites. Finally, any attempt to deduce NEOs' properties from those of meteorites needs a reliable mean to associate a given asteroid or asteroid class to a given meteorite or meteorite group. In this domain, progress has been made recently (e.g. the probable association of the ordinary chondrites with S-complex asteroids) and more can be hoped for in the near future.

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