Chapter 2

Cosmogenic ages of meteorites

2.1 Introduction

It is probably fair to say that the study of meteorites has contributed more to our understanding of the formation of the solar system than any other pursuit (except the basics of celestial mechanics, of course). The amount of information that can be gleaned by the study of noble gases, radioactive isotopes, and their decay products is so great that it would be impossible to attempt to review the entire topic here. I shall concentrate on those studies which provide information that is relevant to the problem of the delivery of the lunar and martian meteorites, but even with this restriction, the body of literature is very large.

Radiometric studies of meteorites (Reedy *et al.* 1983, Vogt *et al.* 1990) allow several quantities to be determined: (1) the crystallization age, *i.e.,* the time at which the various isotopes were locked into a closed system (*e.g.,* the parent rock cools below its molten phase); (2) the 2π depth, which measures how far below the surface of the parent body the meteorite resided before launch; (3) the 2π age $t_{2\pi}$, which is the duration of this surface exposure; (4) the 4π age $t_{4\pi}$ measures the amount of time which the meteorite spent in space exposed to an isotropic flux of cosmic rays; and (5) the terrestrial age $t_{\oplus}$, which is the duration of residence on the Earth's surface before the meteorite was discovered. The terms 2π and 4π refer to the solid angle subtended by the flux of incoming particles that produce the isotopes of interest; objects sitting on the surface of a large body are only exposed to flux coming from 2π steradians. Knowledge of these quantities will allow us to constrain theories about the origin and transfer of the meteorites of interest. The last four of these quantities are dependent on the fact that objects in interplanetary space are exposed to a continuous flux of energetic cosmic rays, whereas the first is in some ways more straightforward. Therefore we will discuss isochron ages first, and cosmic ray exposure methods later.

2.2 Crystallization ages

At the instant that a rock cools and becomes isolated chemically from its surroundings, the radioactive system becomes closed. This results in isotopic ratios involving radionuclides to change with time. If this age can be determined, it is called the *crystallization age* of the object.

To be specific, suppose that a daughter nuclide D is produced in the radioactive decay of a parent nucleus P. Then the rate of change of the concentration (number of atoms per unit volume) of the parent nucleus is given by

$$\frac{dP}{dt} = -\lambda P \qquad \Longrightarrow \qquad P(t) = P_o\, e^{-\lambda t}, \tag{2.1}$$

where P_o is the parent's initial concentration and $\lambda = \ln 2/\tau$ with τ being the usual half-life. Since the daughter isotope may be present in some initial concentration D_o, we write

$$\begin{aligned} D(t) &= D_o + [P_o - P(t)] \\ &= D_o + P(t)\,[\frac{P_o}{P(t)} - 1] \\ &= D_o + P(t)\,[e^{\lambda t} - 1]. \end{aligned} \tag{2.2}$$

This is the fundamental equation of radioactive age determination for a closed system. Since the concentration of an element is difficult to measure, in practice the ratio of concentrations with a stable isotope (usually of the daughter nucleus) is observed, these ratios being determined by mass spectrometry (Wasson 1985). If D_r denotes the reference isotope, then we have

$$\frac{D}{D_r} = \left(\frac{D}{D_r}\right)_o + \frac{P}{D_r}(e^{\lambda t} - 1). \tag{2.3}$$

If different minerals of the same rock having different concentrations of the radioactive parent are sampled, then this equation predicts that the parent and daughter ratios should describe a straight line, called an isochron. The ordinate intercept $(D/D_r)_o$ gives the initial concentration of the daughter isotope, and the slope of this isochron can be used to calculate the time since the system was closed.

An example of this is shown in Fig. 2.1, for the case of the Nakhla meteorite. Here the isotopes used are the very important rubidium-strontium $^{87}\mathrm{Rb} \rightarrow {}^{87}\mathrm{Sr}$ pair, with a 49 Gyr half-life, and $^{86}\mathrm{Sr}$ is used as the stable reference isotope. This is the same method used to date the formation age of almost all of the chondrites at 4.57 ± 0.03 Gyr, which is generally accepted as the time of formation of the solar system. The intercept of this plot shows that (since this corresponds to no initial radioactive Rb present) the initial $^{87}\mathrm{Sr}$ / $^{86}\mathrm{Sr}$ ratio was about 0.7. The

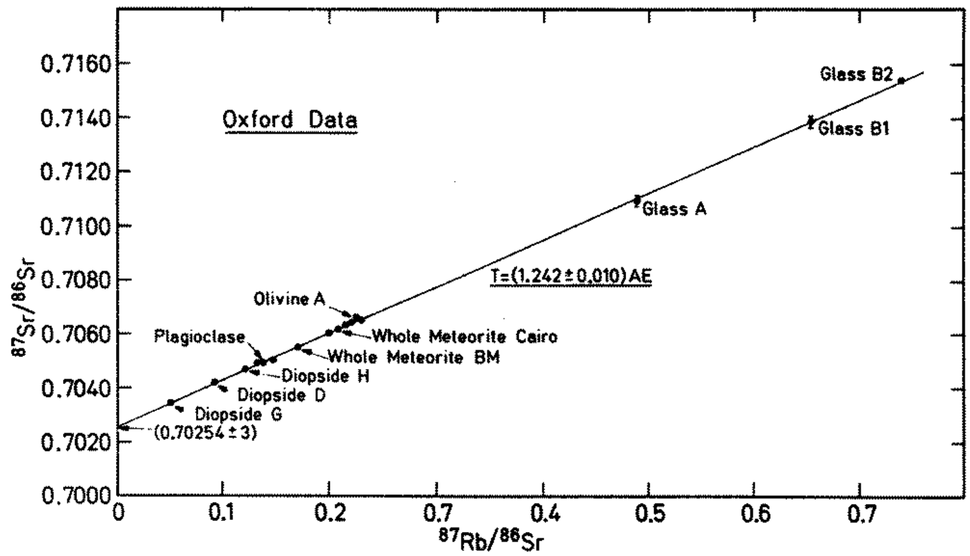

Figure 2.1 An internal isochron for the Nakhla meteorite, using Rb-Sr technique. The large spread in the ^{87}Rb / ^{86}Sr values available at various sites in the meteorite allows an isochron age of 1.24 Gyr to be determined. Figure from Kirsten (1978).

isochron age of Nakhla is 1.24 Gyr, which is surprisingly low, since it implies relatively recent geologic activity.

The crystallization ages of the most lunar meteorites are all very old, varying from 3.9 to 4.3 Gyr (Eugster and Niedermann 1988, Warren and Kallemeyn 1993). This is basically the same as the lunar crust, which remained active for approximately 1 Gyr. Mare basalts can be as young as 3.1 Gyr (Taylor *et al.* 1991); the two mare meteorities Asuka-881757 and Y-793169 have crystallization ages of 3.1 and 4.2 respectively, which is within the normal range although the latter is somewhat high (Eugster *et al.* 1996).

The martian meteorites have 3 groups of crystallization ages, determined from both Rb-Sr and Sm-Nd (Samarium-Neodymium) methods (Nyquist *et al.* 1995). The shergottites (Shergotty, Zagami, EET79001, ALH77005) are dated at $\sim$ 180 Myr, Chassigny and the Nakhlites (Nakhla, Lafayette, and Governador Valadares) at $\sim$ 1.3 Gyr, and ALH84001 at the surprisingly high age of $\sim$ 4.5 Gyr (Jagoutz *et al.* 1994). This last meteorite has caused considerable excitement since it would seem to represent a very ancient piece of the martian crust, in which evidence of a wet early phase for Mars is being sought. The very young crystallization ages of the shergottites may not be crystallization ages at all if a shock event is capable of resetting the Rb-Sr clock; it is possible that they also crystallized $\sim$ 1.3 Gyr ago, but this is in considerable debate (McSween 1995;

Treiman 1995).

The crystallization ages of the martian meteorites are young even for Mars; in fact, terrain younger than 1.3 Gyr probably comprises only 10-15% of the martian surface (Nyquist 1995). Why are 11 of the 12 martian meteorites apparently from this terrain that occupies such a small fraction of the surface? Warren (1994) suggests that the older highland terrain has been pervasively weathered so that almost no intact rocks remain in the upper region of the oldest martian crust. Thus impacts into highland terrain are unable to launch coherent meteoroids. Clearly the existence of ALH84001 must be allowed for if this model is correct.

2.3 Cosmic ray exposure ages

2.3.1 General considerations

The interplanetary flux of cosmic rays causes the production of cosmogenic nuclides (Reedy *et al.* 1983, Vogt *et al.* 1990) in objects that are not shielded by the equivalent of several meters of rock (we say 'equivalent' because, for example, the Earth's atmosphere is also sufficient to shield the surface, decreasing the production rates of radioisotopes by 2–3 orders of magnitude). The isotopes produced can be then studied to determine the duration of the exposure. The cosmic rays of interest for our purposes consist of two types: solar and galactic cosmic rays (SCRs and GCRs), the numbers of which drop off as approximately a power law in the energy E according to $dN/dE \propto E^{-\gamma}$ (Caffee 1988). Both types consist primarily of protons and some alpha particles, with $\simeq 1\%$ heavier nuclei. SCRs of energies about a few MeV can induce nuclear reactions, but only in the upper few mm of matter; the rarer 100 Mev SCRs can penetrate slightly deeper. The GCRs have typical energies of 0.1–10 Gev and can penetrate to a depth of several meters (the flux of primaries decreasing exponentially with depth in the target), causing nuclear reactions not only directly but also through a shower of secondary neutrons that is produced during their passage. Interestingly, nuclide production routes that depend on these secondary neutrons become *more* efficient with depth due to the lack of secondaries near the surface, at least until one gets deep enough in the target that the primaries are screened out (see Englert and Herr 1980 for an example).

The flux of cosmic particles produces the following effects: (1) the implantation of low energy SCRs in the upper 0.1 μm of the target, (2) damage to mineral lattices (mostly in the upper 100 μm) which can be detected under a microscope, and (3) nuclear reactions which transform atoms in the target into new ones (which are then known as cosmogenic nuclides). The most interesting of the created isotopes are noble gases and radioisotopes, both of which are detectable in the sample. Generally the longer the target was subjected to the cosmic ray flux, the greater the concentration of the isotope. Noble gas concentrations (determined as ratios) can be measured by heating samples of the meteorite and observing the escape of the gas (Pal *et al.* 1986, Ott 1988). Activities of radioisotopes can be measured directly (usually in units of disintegrations per minute per kilogram of the sample, denoted dpm/kg) in a nondestructive fashion. All these techniques were refined after the Apollo and Luna missions brought back immense numbers of lunar samples for study.

Clearly the utility of this method depends on knowing the cosmic ray flux over the period of interest, in particular whether or not the flux has been constant. The numbers of both SCRs and GCRs vary with solar activity due to changing SCR production and the fact that the solar magnetic field can shield the inner solar system from GCRs. Over the very short term, the 11-year solar cycle modulates CR activity by changing the rate of solar flares; higher solar magnetic activity prevents the less energetic GCR primaries from reaching the inner solar system. We are more interested in the flux averaged over longer time scales, and in general the data points to the flux being constant in time, although there is some evidence for higher SCR activity in the last hundred thousand years when compared to the last million years. (Reedy *et al.* 1983). This variation is not of great concern since the shallow SCR record is usually removed during the ablation of the meteorite during its atmospheric passage, leaving only the cosmogenic nuclides produced by GCRs available for study. Most cosmogenic nuclides are created by the more energetic GCRs in any case, the flux of which seems to have varied by less than 25–50% in the last few Myr (Reedy *et al.* 1983). A constant flux is assumed by almost all workers in the field.

Cosmogenic *radio*nuclides (as opposed to noble gases) do not continuously accumulate since they are simultaneously decaying. The equation governing the change in their number N (per kg) is $dN/dt = P - \lambda N$, where the first term P is the production rate (events/min/kg) and the second gives the number of decays. Since

$$N(t) = \frac{P}{\lambda}\left(1 - e^{-\lambda t}\right)$$

the activity $A(t)$ (in dpm/kg) that would be measured is

$$A(t) = \lambda N(t) = P(1 - e^{-\lambda t}). \tag{2.4}$$

This produces the expected behavior that, after several half-lives of the isotope, the rate of disintegrations will reach equilibrium with the production rate P, which is thus called the *saturation activity.*

The production rates of the various atomic species depend on the cosmic ray flux, and the composition of the target. In general, the formation rate of a given nuclide x at depth d inside a meteoroid of nominal radius R is given by (Vogt *et al.* 1990)

$$P_x(d,R) = \sum_j N_j \sum_k \int_0^\infty \sigma_{xjk}(E)\ \phi_k(E,d,R)\ dE\ , \tag{2.5}$$

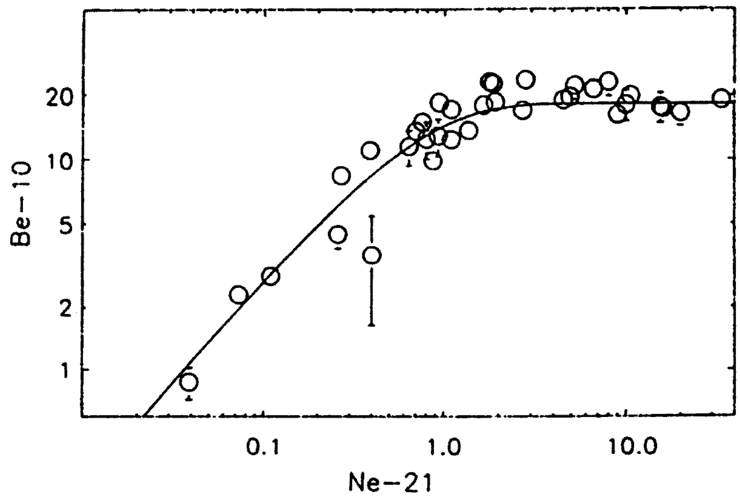

Figure 2.2 ^{10}Be activities (dpm/kg) versus ^{21}Ne content (volume per gram, measured in 10^{-8}cm^3 STP g^{-1}) for stony meteorites. From Vogt *et al.* (1990).

where N_j represents the concentration of target element j, σ_{xjk} is the cross section for the production of x from element j by particle type k at energy E, and ϕ is the flux of these particles. As can be imagined, determining these cross-sections experimentally is a time consuming task. The production rate's dependence on the target's composition produces formulae such as

$$P_{10} = 0.30\mathrm{O} + 0.26(\mathrm{Mg} + 0.8\mathrm{Al} + 0.65\mathrm{Si}) + 0.12\mathrm{S} + 0.10\mathrm{Ca} + 0.07(\mathrm{Fe} + \mathrm{Ni}) \qquad (2.6)$$

for the saturation activity of ^{10}Be in stones, where the elemental names refer to the fractional abundance (by weight) of that element in the sample (Pal *et al.* 1986).

If one wishes to confine one's attention to a particular meteorite type (so that the compositions are similar), then the production rate can be measured in a much simpler way. Consider that stable isotopes (such as noble gases) are also being produced at a rate P_s, but that the concentration S of these nuclides will increase linearly since they do not decay; that is:

$$S = P_s\, t\ . \qquad (2.7)$$

If the activity of radionuclide is plotted against the concentration of the stable isotope for a variety of meteorites of similar composition, then a plot like Fig. 2.2 results, where the ^{10}Be activity is plotted against the ^{21}Ne content (which thus serves as a clock) and shows the ^{10}Be activity reaching saturation. Since Eq. (2.7) can be used to eliminate the time in Eq. (2.4) to yield $^{10}\mathrm{Be} = P_{10}[1 - \exp(-\lambda\ ^{21}\mathrm{Ne}/P_{21})]$, a two parameter fit to the data yields *both* production rates.

The production rates obtained by a variety of methods agree within 10 or 20%. The rates used in analyses of the lunar and martian meteorites are typically those produced in stony meteorites (specifically chondrites), with minor modification

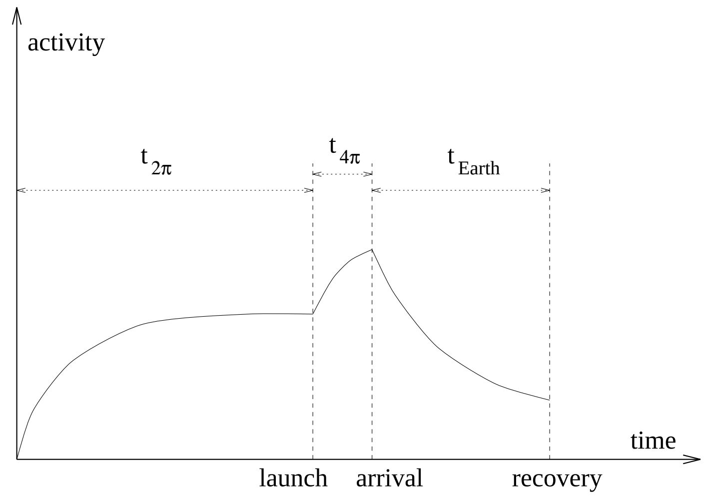

Figure 2.3 Schematic representation of the time evolution of the activity of a radionuclide. In this example the meteoroid is close enough to the surface for many half-lives so that the 2π saturation activity is reached. During the short transfer ($t_{4\pi}$) the new saturation level is not obtained, and then during Earth residence time ($t_{\rm Earth} = t_{\oplus}$) the activity decays away continuously.

(Pal *et al.* 1986). A final correction needs to be done to allow for the effects of shielding within the meteorite itself. Measurement of the noble gas concentration ratio ^{3}He / ^{21}Ne (which is diagnostic of the shielding depth) combined with estimates of the pre-atmospheric size yield a corrected production rate which is then used in the age calculation.

The overall errors in the CRE ages, as judged by the results determined using different isotopes, appear to be roughly 10–30%. This accuracy should be sufficient for our purposes. The overall problem now becomes one of disentangling where, and for how long, the various irradiations and shieldings occurred. For the lunar and martian meteorites, this consists of three possible phases: (1) irradiation on the surface of the parent body for $t_{2\pi}$, where the depth of burial is important, (2) irradiation in space ($t_{4\pi}$), and (3) the Earth residence time $t_{\oplus}$, during which the meteorite is effectively shielded. A current measured activity A could unfortunately result from several different exposure histories, which satisfy

$$A = [\ P_{pb}(1 - e^{-\lambda t_{2\pi}})e^{-\lambda t_{4\pi}} + P_{4\pi}(1 - e^{-\lambda t_{4\pi}})\ ]\ e^{-\lambda t_{\oplus}}. \tag{2.8}$$

Here the production rate on the parent body P_{pb} accumulates for $t_{2\pi}$, and then this contribution will decay away while the meteoroid is in space for $t_{4\pi}$. However, the 4π production rate $P_{4\pi}$ is always greater and so the activity will increase until the object reaches the Earth, whereupon the final exponential factor in Eq. (2.8) indicates both contributions continue to decay on the Earth. This is illustrated schematically in Fig. 2.3, which is a reasonable representation of the history of a lunar meteorite. In contrast, martian meteorites have no surface exposure (Sec. 2.3.4), and so their activities would be zero at the moment of launch. Since there are five unknowns (if we include the production rates), many different histories could lead the same final activity. For example, in the figure the same final activity could result by having an instantaneous transfer ($t_{4\pi} = 0$) followed by a shorter Earth residence time. Fortunately, by measuring more than one isotope (ideally five), this ambiguity can be removed. It can be difficult to find enough radioisotopes with half-lives in the appropriate regimes; four is usually sufficient since $P_{4\pi}$ can be estimated (Nishiizumi *et al.* 1991). We now discuss results for these various portions of the exposure histories, best estimates of which are shown in Table 2.1.

2.3.2 4π exposure ages

The 4π ages will be our chief constraint since we are primarily interested in how much time elapsed between launch and Earth impact. It is apparent from an examination of Table 2.1 is that the average lunar meteorite was delivered to the Earth in a far shorter time than the average (or indeed, almost any) martian meteorite. Over half of the lunar meteorites have transfer durations of less than 100 kyr, a fact that will serve as an important constraint later. Also observe that the martian meteorites appear to fall into three CRE age groups: 0.7 Myr (1 object), 3 Myr (5 objects), and about 10 ± 1 Myr (4 objects), with ALH84001 is almost certainly independent of this last group based on CRE and compositional data (Bogard 1995). We will explore what these transfer data imply for our dynamical model in the following chapters.

2.3.3 Earth residence and blast-off times

Except for the lunar meteorite Calcalong Creek and the martian falls and finds, all of the meteorites with measured values of $t_\oplus$ were found in Antarctica and have residence times consistent with what we know are typical ages for the collection areas (Nishiizumi *et al.* 1989, Vogt 1990); the oldest Antarctic meteorites are about 1 Myr old, but in some collection areas the average is considerably younger (for example, < 0.1 Myr for the Yamato Mountains). When performing this analysis one wants to use a radioisotope with a half-life on the same order as the Earth residence time, and so ^{36}Cl ($\tau_{1/2} = 0.3$ Myr) and ^{14}C (5740 yr) are good choices (Nishiizumi 1986).

Even though $t_\oplus$ is rather short for most meteorites, since many lunar meteorites have such small 4π ages it is important to know $t_\oplus$ since the launch time (or *blast-off time*) was $t_{bo} = t_{4\pi} + t_\oplus$ years ago. Two meteorites that do not share similar blast-off times cannot be source-crater paired.

Table 2.1 Properties of the lunar and martian meteorites, summarized from Warren (1994); a typo in the mass of EET79001 is corrected here. The abbreviation Y = Yamato is used. The martian meteorites that are labeled falls have Earth residence times of effectively zero, and those labelled finds likely have very young $t_\oplus$ (negligible compared to $t_{4\pi}$). The martian meteorites all have $t_{2\pi} = 0$ due to their deep burial (> several meters). The parenthesized 4π ages for some of the lunar meteorites are thermoluminescence transit ages (see ref. 3). REFERENCES: 1. Warren 1994, 2. Nishiizumi *et al.* 1991, 3. Benoit *et al.* 1996, 4. Nishiizumi 1996a, 5. Swindle *et al.* 1995, 6. Eugster *et al.* 1996, 7. Nishiizumi 1996b, 8. Thalmann *et al.* 1996.

Lunar Meteorite	mass (kg)	$t_{4\pi}$ (Myr)	$t_\oplus$ (Myr)	$t_{2\pi}$ (Myr)	depth (g/cm^2)	ref.
Y-793274	0.009	< 0.02	$<.02$–$.06$	>300	140–180	1
MAC88104/5	0.724	~ 0.04	$0.23\pm.02$	630	360–400	1
ALH81005	0.031	<0.05(0.01)	0.01–.02	580	150–180	2,3
QUE94281	0.023	$\ll$0.05(.001)	0.1–0.2	>100	270–320	4
EET87521	0.031	<0.07(<.001)	<0.06	<0.2	560–590	1,3
Y-791197	0.052	$\ll$0.1(<.019)	$0.06\pm.03$	910	4–8	1
QUE93069/94069	0.024	$0.15\pm.02$	<0.015	1000	10–50	4,8
Calcalong Creek	0.019	0.2	<0.07	>200	~ 40	5
Asuka-881757	0.442	0.9 ± 0.1	<0.05	0.8	>1000	4,8
Y-793169	0.006	1.1 ± 0.2	<0.05	50 ± 10	>500	4,8
Y-82192,3/86032	0.712	9 ± 2	$\sim$0.09	none	>1000	1

Martian Meteorite	mass (kg)	4π CRE (Myr)	$t_\oplus$ (Myr)	$t_{2\pi}$ (Myr)	depth (m)	ref.
EET79001	7.94	0.6	0.012	–	> 2	6,7
QUE94201	0.012	2.6	0.3	–	> 2	7
Zagami	18	2.6	fall	–	> 2	6
Shergotty	4	2.8	fall	–	> 2	6
ALH77005	0.48	3.4	0.2	–	> 2	6
LEW88516	0.013	3.6	0.025	–	> 2	6
Chassigny	4	9.9	fall	–	> 2	6
Governador Valadares	0.16	10.5	find	–	> 2	6
Nakhla	40	10.5	fall	–	> 2	6
Lafayette	0.80	10.5	find	–	> 2	6
ALH84001	1.93	$\approx$15	0.011	–	> 2	6
Y-793605	0.018	?	?	?	?	–

The fact that roughly half of the martian meteorites were found outside of Antarctica (many as large falls) probably requires explanation. Is the lack of large lunar meteoroids in historical times just a statistical fluke, or does it indicate something about the launch process?

2.3.4 2π exposure and surface depths

The 2π exposure ages are also obtained from an analysis of the activities of multiple nuclides. Looking for this kind of 'first stage' activity is often not done for meteorites, which are often assumed to have simple '1–stage' exposure histories (that is, the meteorite is liberated from a completely shielded location and then transferred to the Earth with $t_{4\pi} \gg t_{\oplus}$). This is because the liberation event is often assumed to be the catastrophic disruption of a large parent asteroid, which suddenly exposes massive numbers of meteoroids to cosmic rays. This assumption has been questioned by Wetherill (1980) who believes that complex histories should be more the norm. But in the case of the lunar and martian meteorites, near-surface launch from an impact event immediately suggests that the objects may have resided on or near the surface of the parent body.

The results for the lunar meteorites show in almost all cases that the objects spent several hundred Myr to 1 Gyr within the active layer of the upper few meters of the surface (Nishiizumi *et al.* 1991, Warren 1994). This lunar depth information is obtained by using profiles for the saturation activity taken from Apollo drill core samples within which the activities are known as a function of depth (Carrier *et al.* 1991). Once it is known that the 2π age is long, the drill core activities can be used to determine the depth at which the meteoroid resided before launch. The depth is quoted as an integrated mass depth d_m (in g/cm^2) defined by

$$d_m = \int_0^X \rho(z)\; dz \quad , \tag{2.9}$$

where X is the depth of burial (cm) of the meteorite. Thus, in a uniform surface layer the real depth of burial is therefore just the mass depth divided by the mean density.

A complication is presented in that there exists the possibility that the meteoroid could have been exposed for some time and then buried (by a nearby impact turning over the regolith for example). In such a case the activity produced by the previous exposure would begin to decay away. This confusion can be eliminated by examining stable isotopes (mostly noble gases) that do not disappear with time and which thus measure the integrated surface residence age. In any case, since the interval between a burial and re-exposure is likely to be long compared to the half-life, the 2π age and depth measures the time and depth since the last emplacement in the near-surface layer.

On the basis of this and noble gas concentrations, the conclusion is that the martian meteoroids were all launched from completly shielded locations, *i.e.,* at least several meters below the surface, implying $t_{2\pi} = 0$ (McSween 1995). The deep pre-launch residence of the martian meteorites likely provides clues as to the ejection mechanism, and this difference has prompted speculation that the launch process may differ in some manner from that for the lunar meteorites (Ott 1988, Warren 1994).