AGES AND GEOLOGIC HISTORIES OF MARTIAN METEORITES

L.E. NYQUIST[1], D.D. BOGARD[1], C.-Y. SHIH[2], A. GRESHAKE[3], D. STÖFFLER[3] and O. EUGSTER[4]

[1]*SN/Planetary Sciences, NASA Johnson Space Center, 2101 NASA Road 1, Houston, TX 77058-3696, USA*

[2]*Basic and Applied Research Department, Lockheed-Martin Space Operations, 2400 NASA Road 1, Houston, TX 77058-3696, USA*

[3]*Institut für Mineralogie, Museum für Naturkunde, Invalidenstr. 43, D-10115 Berlin, Germany*

[4]*Physikalisches Institut, University of Bern, Sidlerstr. 5, CH-3012 Bern, Switzerland*

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Abstract. We review the radiometric ages of the 16 currently known Martian meteorites, classified as 11 shergottites (8 basaltic and 3 lherzolitic), 3 nakhlites (clinopyroxenites), Chassigny (a dunite), and the orthopyroxenite ALH84001. The basaltic shergottites represent surface lava flows, the others magmas that solidified at depth. Shock effects correlate with these compositional types, and, in each case, they can be attributed to a single shock event, most likely the meteorite's ejection from Mars. Peak pressures in the range $15 - 45$ GPa appear to be a "launch window": shergottites experienced $\sim 30 - 45$ GPa, nakhlites $\sim 20 \pm 5$ GPa, Chassigny ~ 35 GPa, and ALH84001 $\sim 35 - 40$ GPa. Two meteorites, lherzolitic shergottite Y-793605 and orthopyroxenite ALH84001, are monomict breccias, indicating a two-phase shock history *in toto*: monomict brecciation at depth in a first impact and later shock metamorphism in a second impact, probably the ejection event.

Crystallization ages of shergottites show only two pronounced groups designated S_1 ($\sim$175 Myr), including 4 of 6 dated basalts and all 3 lherzolites, and S_2 ($330 - 475$ Myr), including two basaltic shergottites and probably a third according to preliminary data. Ejection ages of shergottites, defined as the sum of their cosmic ray exposure ages and their terrestrial residence ages, range from the oldest ($\sim$20 Myr) to the youngest ($\sim$0.7 Myr) values for Martian meteorites. Five groups are distinguished and designated S_{Dho} (one basalt, $\sim$20 Myr), S_L (two lherzolites of overlapping ejection ages, 3.94 ± 0.40 Myr and 4.70 ± 0.50 Myr), S (four basalts and one lherzolite, $\sim 2.7 - 3.1$ Myr), S_{DaG} (two basalts, $\sim$1.25 Myr), and S_E (the youngest basalt, 0.73 ± 0.15 Myr). Consequently, crystallization age group S_1 includes ejection age groups S_L, S_E and 4 of the 5 members of S, whereas S_2 includes the remaining member of S and one of the two members of S_{DaG}. Shock effects are different for basalts and lherzolites in group S/S_1. Similarities to the dated meteorite DaG476 suggest that the two shergottites that are not dated yet belong to group S_2. Whether or not S_2 is a single group is unclear at present. If crystallization age group S_1 represents a single ejection event, pre-exposure on the Martian surface is required to account for ejection ages of S_L that are greater than ejection ages of S, whereas secondary breakup in space is required to account for ejection ages of S_E less than those of S. Because one member of crystallization age group S_2 belongs to ejection group S, the maximum number of shergottite ejection events is 6, whereas the minimum number is 2.

Crystallization ages of nakhlites and Chassigny are concordant at $\sim$1.3 Gyr. These meteorites also have concordant ejection ages, i.e., they were ejected together in a single event (NC). Shock effects vary within group NC between the nakhlites and Chassigny.

The orthopyroxenite ALH84001 is characterized by the oldest crystallization age of $\sim$4.5 Gyr. Its secondary carbonates are $\sim$3.9 Gyr old, an age corresponding to the time of Ar-outgassing from silicates. Carbonate formation appears to have coincided with impact metamorphism, either directly, or indirectly, perhaps via precipitation from a transient impact crater lake.

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The crystallization age and the ejection age of ALH84001, the second oldest ejection age at 15.0 ± 0.8 Myr, give evidence for another ejection event (O). Consequently, the total number of ejection events for the 16 Martian meteorites lies in the range 4 − 8.

The Martian meteorites indicate that Martian magmatism has been active over most of Martian geologic history, in agreement with the inferred very young ages of flood basalt flows observed in Elysium and Amazonis Planitia with the Mars Orbital Camera (MOC) on the Mars Global Surveyor (MGS). The provenance of the youngest meteorites must be found among the youngest volcanic surfaces on Mars, i.e., in the Tharsis, Amazonis, and Elysium regions.

Keywords: shock effects, crystallization ages, cosmic ray exposure ages, ejection ages, provenance

1. Introduction

The clan of Martian meteorites, formerly called SNCs after *S*hergotty, *N*akhla and *C*hassigny, now consists of 16 unpaired meteorites of magmatic origin (basalts and ultramafic cumulates). Generally young crystallization ages (with the exception of one pyroxenite), characteristic isotopic compositions of C, N, O, and noble gases, as well as distinct major and trace element concentrations, distinguish them from all other differentiated meteorites (McSween, 1994; Clayton and Mayeda, 1996; Dreibus and Wänke, 1987; Wänke and Dreibus, 1988; Wänke, 1991). The most convincing evidence for the Martian origin of these rocks is given by isotopic measurements of trapped gases in shock-melted glass of shergottites. It was found that the isotopic composition of these gases is indistinguishable from Martian atmosphere within the measurement errors of the mass spectrometer on board the Viking lander (e.g., Bogard and Johnson, 1983; Becker and Pepin, 1984; Swindle *et al.*, 1986; Marti *et al.*, 1995).

Those who have followed the development of the hypothesis of the Martian origin of these meteorites will recognize that much of the early information about them was obtained in attempts to verify that hypothesis. Once the determination of probable Martian origin had been made, observations about the meteorites could be generalized to become probable observations about Mars. Efforts to do so are hampered to variable degrees by lack of knowledge of the geologic settings from whence the meteorites came. This is also true of the radiometric age data.

The young radiometric ages of shergottites, first observed for the type example, Shergotty, by Geiss and Hess (1958), were among the first lines of evidence cited in support of their origin on a planetary-sized body, probably Mars (Nyquist *et al.*, 1979b; Wasson and Wetherill, 1979). However, some characteristics of the analytical data seemed to violate criteria developed for unambiguous interpretation of radiometric ages as igneous crystallizaton ages. In particular, the Rb-Sr and ^{39}Ar-^{40}Ar ages were discordant (Nyquist *et al.*, 1979a; Bogard *et al.*, 1979). Moreover, the Rb-Sr data showed considerable "scatter" about the best fit isochron. Thus, it did not initially appear possible to interpret the ∼165 Myr Rb-Sr age of Shergotty, for example, as the crystallization age, particularly, since the ^{39}Ar-^{40}Ar age of a plagioclase separate was older at ~250 Myr. Considerable experience with dating lunar samples had shown that although ^{39}Ar-^{40}Ar ages of bulk samples could be biased low due to diffusive loss of ^{40}Ar, ^{39}Ar-^{40}Ar ages of plagioclase separates generally gave reliable crystallization ages. Thus, the Rb-Sr age was initially interpreted as likely reflecting the time of a thermal metamorphism. Because the shergottites were more highly shocked than almost all other meteorites or lunar samples, initial attention was given to post-shock thermal metamorphism as the agent for resetting both types of ages (Nyquist *et al.*, 1979a; Bogard *et al.*, 1979). It has since been established that neither post-shock thermal metamorphism nor shock transformations of mineral phases are adequate to reset the Rb-Sr isotopic system. Thus, the *possibility* of shock resetting of the radiometric ages appears to have been a false lead.

Radiometric ages determined by other methods often were discordant also, and contributed to the confusion about the ages of shergottites. Here, it is sufficient to note that isotopic heterogeneities occurring within the rocks over distances of centimeters appear to complicate the isotopic data. For example, the existence of heterogeneity in initial $^{87}Sr/^{86}Sr$ between different samples of Zagami was shown by Nyquist *et al.* (1995). Papanastassiou and Wasserburg (1974) had observed a similar heterogeneity in initial $^{87}Sr/^{86}Sr$ for Nakhla much earlier. They took their observation as evidence that the Rb-Sr age of 1.30 ± 0.02 Gyr that they determined separately for two samples of Nakhla was an age of metamorphism rather than an igneous crystallization age. (Here, as elsewhere in this paper, we use the value of the ^{87}Rb decay constant recommended by Minster *et al.* (1982); i.e., $\lambda_{87} = 1.402 \times 10^{-11} yr^{-1}$). Gale *et al.* (1975) obtained apparently well-defined, concordant, isochrons for two additional samples of Nakhla. They interpreted the isochron age for their total data set, 1.23 ± 0.01 Gyr, as the age of igneous crystallization. The present authors agree with that interpretation, but the issue of possible sample heterogeneity exists. The two samples studied by Papanastassiou and Wasserburg (1974) were obtained from separate sources, and weighed 0.7 g and 2.3 g, respectively (D. Papanastassiou, personal communication). The samples studied by Gale *et al.* (1975), also from two different sources, weighed 13 g and 18 g, respectively. It is possible that the larger samples used by Gale *et al.* (1975) effectively averaged out isotopic heterogeneity that could exist in Nakhla over small distances.

In spite of uncertainties in interpreting the isochron data, the early isotopic data of both Nakhla and Shergotty showed unambiguously that they were "young" by meteorite standards. For Nakhla, the evidence was Rb-Sr model ages in the range of $2.5 - 3.6$ Gyr, calculated relative to the initial $^{87}Sr/^{86}Sr$ of the solar system (Papanastassiou and Wasserburg, 1974; Gale *et al.*, 1975). These model ages provide a strict upper limit to the formation age of Nakhla. The Rb-Sr model ages of shergottites are ~4.5 Gyr, indicative of the time of chemical differentiation of their parent body. Nyquist *et al.* (1979b) found the Sm-Nd model ages of three shergottites relative to initial $^{143}Nd/^{144}Nd$ of the solar system to be only $\sim 2.8 - 3.6$ Gyr, however. These Sm-Nd model ages also provide a strict upper limit to the formation ages of the meteorites, and they are especially remarkable in that they imply significant fractionation of the Sm/Nd ratios of the meteorites from chondritic values, which normally is not achieved in "simple" magmatic processes. Thus, these young model ages clearly pointed to "late" and complex magmatic activity on the parent body of the shergottites.

Another feature of the shergottites that makes them somewhat unusual among stony meteorites is their relatively young cosmic ray exposure (CRE) ages. Such ages are determined from the accumulation of nuclides spalled from target nuclides in the meteorites by high energy cosmic ray interactions. Spallation nuclides that are stable against radioactive decay accumulate continually at production rates that are a function of the chemical composition of the meteorite and of changes in the energy spectrum of primary and secondary cosmic rays as a function of the size of the meteoroid and the depth of the meteorite within it. Spallation radionuclides that undergo radioactive decay accumulate only to equilibrium levels at which they decay at the same rate at which they are produced. The production rates of stable nuclides can be determined either theoretically using nuclear cross section data, or empirically from the equilibrium activities of closely related radionuclides for which the ratio of the production rate to that of the stable nuclide is known. Determination of stable noble gas CRE ages, for example, require mass spectrometric techniques similar to those required for ^{39}Ar-^{40}Ar age dating, so stable noble gas CRE ages often are determined in the same investigations as those in which ^{39}Ar-^{40}Ar ages are determined. An early study of four shergottites, Shergotty, Zagami, ALH77005, and EET79001 (Bogard *et al.*, 1984) showed that, whereas three of them had very similar apparent exposure ages of $\sim$2–3 Myr, one (EET79001) had a much lower apparent exposure age, $<$ 1 Myr. One possible explanation was that the actual exposure age of EET79001 was the same as that of the others, but that the production rate of stable nuclides was lower in EET79001 because it was located deeper in a large meteoroid, where the effective flux of primary cosmic rays was attenuated by about a factor of 3–4. A variation of this scenario is that EET79001 was initially part of a very large meteoroid, and was essentially totally shielded from cosmic rays for most of its lifetime in space. With acquisition of additional data for both stable and radionuclides, both of these explanations have fallen into disfavor. The currently accepted interpretation of the spallogenic nuclide data is that exposure to cosmic rays was initiated by excavation of the meteorites from depths on Mars that were completely shielded from the effects of cosmic rays, and thus that the CRE ages plus the time the meteorites have been on Earth give directly the time since they were ejected from Mars (*cf.* Eugster *et al.*, 1997b). However, the young CRE age of EET79001 presents a problem when viewed in the context of its crystallization age, and its apparent relationship to the other shergottites.

Thus, the earlier interpretation of the Rb-Sr and ^{39}Ar-^{40}Ar ages for shergottites of $\sim$180 Myr and $\sim$250 Myr, respectively, as giving the time of shock metamorphism during their ejection from Mars as large blocks to be broken up in later secondary collisions (*cf.* Shih *et al.*, 1982) is no longer favored by the majority of of meteoriticists. Also, a steadily increasing number of Sm-Nd ages have been obtained that are concordant with the Rb-Sr age for the same meteorite. Furthermore, ^{39}Ar-^{40}Ar ages can be explained by the presence of excess, non-radiogenic ^{40}Ar from a variety of sources. These problems have been much less acute for the other types of Martian meteorites, for which the ages are comparatively well defined.

The desire to find possible mechanisms for launching Martian meteorites and for metamorphic resetting of the radiometric ages of shergottites stimulated thorough investigations of their shock-metamorphic features. These shock features of the meteorites provide insight into the environment of the rocks when they were ejected from Mars. Furthermore, the original igneous textures of their minerals provide insight into the environments in which the rocks originally crystallized. For example, mineral textures are influenced by the cooling rate of the rock during crystallization, which in turn is influenced by the thickness of a magma flow. Thus, observable features in the meteorites themselves tell us some things about their geologic setting at key points in their histories.

Because the Martian meteorites are the only samples from Mars currently available for laboratory studies, their properties are of great importance to understand the formation and evolution of our neighboring planet. In this review we summarize their crystallization and cosmic ray exposure ages, compare the shock levels to which they have been exposed, and briefly consider the implications of those data for the meteorites' provenance and for Martian evolution. But, first we describe the mineralogical and geochemical characteristics of the meteorites themselves.

2. Mineralogy, Petrography, and Geochemistry

According to their mineralogical composition and textural characteristics, the Martian meteorites represent igneous rocks of basaltic and ultramafic provenance. They appear to have crystallized either in lava flows as volcanic rocks or in mafic, probably shallow, intrusions as plutonic ultramafic rocks. They are divided into shergottites, consisting of a basaltic and a lherzolitic subgroup, nakhlites (clinopyroxenites), chassignites (dunites), and orthopyroxenites. Chassigny and ALH84001 are the only dunite and orthopyroxenite in the latter two groups (Figure 1). In the following, we describe the mineralogy, petrography, and geochemistry of the various Martian meteorites. For more details, the reader is referred to the review article by McSween (1994) and to the Mars Meteorite Compendium (Meyer, 1998).

2.1. Basaltic Shergottites

The meteorites Shergotty, Zagami, EETA79001, QUE94201, Dar al Gani 476, and Los Angeles, as well as the recently found Dhofar 019 and Sayh al Uhaymir 005, belong to the group of basaltic shergottites. These rocks predominantly consist of augite and pigeonite, typically showing a strong irregular chemical zoning towards an Fe-rich rim, and of plagioclase in the form of shock-induced diaplectic

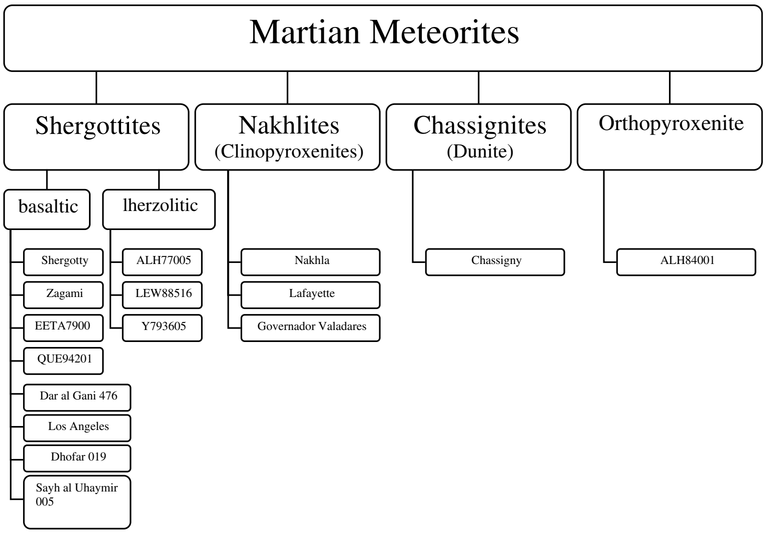

Figure 1. Classification of Martian meteorites (modified after Stephan *et al.*, 1999).

glass (maskelynite). Minor components are pyrrhotite, whitlockite, ilmenite and titanomagnetite (McSween, 1994). Pyroxene often contains small rounded to subrounded melt inclusions of kaersutite, spinel, and sulfides in a Si-rich glassy or microcrystalline groundmass, interpreted as trapped original melt (e.g., Treiman, 1985). EETA79001 contains two basaltic lithologies, termed A and B. The main differences between these units are the small grain size and the occurrence of olivine xenocrysts, orthopyroxene, and chromite in lithology A. Besides formation by simple mixing of basaltic liquids, it was suggested that lithology A represents an impact melt (Mittlefehldt *et al.*, 1997). Among the basaltic shergottites, only Dar al Gani 476 and Sayh al Uhaymir 005 have some similarities to lithology A of EETA79001, containing large xenocrysts of olivine set into a matrix of clinopyroxene and maskelynite (Zipfel *et al.*, 2000; Grossman, 2000). The basaltic shergottites, except QUE94201 and Los Angeles, show a cumulate texture with mostly preferred orientations of the pyroxenes. However, the relatively small grain size of the pyroxenes and the petrographic similarities among the meteorites suggest that the alignment may have occurred by lava flow rather than by accumulation in a subsurface magma chamber (McCoy *et al.*, 1992; McSween, 1994).

Compared to terrestrial basalts, the basaltic shergottites are characterized by a high Fe/(Fe+Mg) ratio and low Al_2O_3 concentrations. All meteorites of this group have complex rare Earth element (REE) patterns with distinct depletions of light REE (LREE) but without a clear Eu anomaly. Except for water, all volatile elements are enriched (Wänke and Dreibus, 1988; Lodders, 1998; Zipfel *et al.*, 2000; Rubin *et al.*, 2000). Initial Sr, Nd, and Pb isotopic compositions are variable among the basaltic shergottites, possibly reflecting different stages of mixing Martian crust with an isotopically homogeneous magma.

2.2. Lherzolitic Shergottites

Three shergottites (ALH77005, LEW88516, and Y-793605) are lherzolitic as they contain $<$10 vol.% plagioclase. They are composed of relatively coarse-grained anhedral to euhedral olivine and chromite enclosed by large orthopyroxene crystals (Harvey *et al.*, 1993; McSween, 1994; Mikouchi and Miyamoto, 1997). Interstices are filled with accessory phases, i.e. maskelynite, pigeonite, augite, and whitlockite. The Fe-Mg-silicates of this suite of ultramafic rocks are much more magnesian than those of the basaltic shergottites, and an observed chemical disequilibrium between coexisting olivine and pyroxene gives evidence for non-linear cooling (Harvey *et al.*, 1993). High concentrations of Fe^{3+} in chromites suggest high oxygen fugacity during crystallization. Similarly to the basaltic shergottites, the lherzolitic ones are depleted in LREE (Dreibus *et al.*, 1982). However, the composition of radiogenic Sr isotopes in ALH77005 and LEW88516 shows that the two rocks crystallized from different magma sources (Borg *et al.*, 1998a, 1998b).

2.3. Nakhlites

The group of nakhlites contains the three clinopyroxenites Nakhla, Lafayette, and Governador Valadares. They consist of Mg-rich augite and Fe-rich olivine set into a microcrystalline groundmass of mostly radiating crystalline plagioclase, which has not been transformed into maskelynite by shock, pigeonite, ferroaugite, titanomagnetite, pyrite, troilite, chlorapatite, and sometimes SiO_2-rich glass. In addition, they contain phyllosilicates (iddingsite) and evaporite mineral assemblages of secondary, but Martian, origin confirming the presence of liquid water on Mars (Gooding *et al.*, 1991; Bridges and Grady, 2000; Swindle *et al.*, 2000).

All three meteorites are cumulates, and lamellar inclusions of augite and magnetite in olivine from Nakhla and Governador Valadares confirm slow cooling under highly oxidizing conditions (Mikouchi and Miyamoto, 1998). In addition, olivine frequently contains small melt inclusions compositionally representing a major and distinct type of Martian magma.

All nakhlites have moderately high contents of volatiles and are enriched in LREE. The almost identical initial Sr and Nd isotopic composition of nakhlites is distinct from that of the shergottites, indicating that the two types of rocks formed from different parent magmas (Nakamura *et al.*, 1982b; McSween, 1994).

2.4. Chassigny

Chassigny is the only dunite among the Martian meteorites, and consists of 90% Fe-rich olivine ($Fa_{\sim 32}$), 5% pyroxene, 2% feldspar ($An_{\sim 20}$; maskelynite), and 3% accessory phases (Floran *et al.*, 1978). High concentrations of Fe^{3+} in chromite as well as lamellar exsolutions in olivine indicate crystallization at high oxygen fugacity and low cooling rates (Floran *et al.*, 1978; Greshake *et al.*, 1998). Its cumulus fabric indicates a fractional crystallization of a mafic magma body.

Chassigny is enriched in LREE but shows no Eu anomaly. While this pattern is distinct from those of nakhlites, excluding a formation from the same magma, the initial Sr isotope compositions of Chassigny and nakhlites are identical (Nakamura *et al.*, 1982a; McSween, 1994).

2.5. ALH84001

ALH84001 is a coarse-grained brecciated orthopyroxenite with a modal composition of 96% orthopyroxene, 2% chromite, 1% plagioclase (maskelynite), and 0.15% phospate. Accessory phases are augite, olivine, pyrite and Fe-Mg-Ca-carbonates (Mittlefehldt, 1994). Texturally, ALH84001 is dominated by up to 6 mm long orthopyroxene crystals joined at 120° triple junctions and poikilitically enclosing Fe^{3+}-rich euhedral chromites (Berkley and Boynton, 1992; Mittlefehldt, 1994; McSween, 1994). Maskelynite and rarely chromite occur interstitially between orthopyroxene. Predominantly along fractures and in cataclastic areas, compositionally strongly zoned carbonates are found, often forming characteristic globules that appear either as concentric spherules or as flat "pancakes" (e.g., McKay *et al.*, 1996). Many carbonates resemble a "bull's-eye" with a center of dolomite-ankerite surrounded by concentric bands of siderite, magnesite, and sulfide, but they come in a multitude of varieties (Scott *et al.*, 1998). High-resolution Scanning Electron Microscope (SEM) images revealed worm-like features in the carbonates. Additionally, the morphologies of some magnetite grains in the carbonates resemble those formed by magnetotactic bacteria, and relatively high concentrations of polycyclic aromatic hydrocarbons (PAHs) have been found. From these observations, McKay *et al.* (1996) concluded early biologic activity was present on Mars. Meanwhile, various other non-biogenic formation mechanisms of the carbonates and magnetite assemblage have been proposed, including impact origin (Harvey and McSween, 1996; Scott *et al.*, 1998; Scott, 1999) and flood-evaporite formation (McSween and Harvey, 1998; Warren, 1998). The high concentrations of PAHs, present in all textural units of ALH84001 (Stephan *et al.*, 1998, 1999), could also be due to terrestrial contamination (Becker *et al.*, 1997; Jull *et al.*, 1998).

ALH84001 is depleted in the LREE and has a negative Eu anomaly. Its very low concentrations of siderophile elements led to the development of a model for the Martian mantle depleted in siderophile elements (Dreibus *et al.*, 1994).

2.6. ENVIRONMENTS OF IGNEOUS CRYSTALLIZATION

2.6.1. *Basaltic Shergottites*

The textures of the basaltic shergottites are consistent with those expected for surface flows of basaltic lava. McCoy *et al.* (1992) suggested that Zagami was the product of a two-stage magmatic history. The first stage occurred in a slowly cooling magma chamber. The presence of amphibole in the cores of pyroxene crystals requires pressures equivalent to depths >7.5 km on Mars. During the second stage, pyroxene crystals were entrained into a magma that either intruded to the near surface and cooled in a relatively thin dike or sill, or extruded to the surface and crystallized in a lava flow >10 m thick. This two-stage scenario is consistent with observations of volcanic constructs and flows in the Tharsis region of Mars.

2.6.2. *Lherzolitic Shergottites*

Detailed investigations of lherzolitic textures revealed a preferred crystallographic orientation of olivine, proving that the lherzolitic shergottites are real cumulates, formed in a plutonic sub-surface environment (Berkley and Keil, 1981; McSween, 1994). Ikeda (1994) suggested that the compositional discontinuities among the four zoning types of chromite in ALH77005 arise from magma mixing in shallow magma reservoirs on Mars. Their crystallization histories, as reconstructed by Harvey *et al.* (1993) and McSween (1994), require varying degrees of prolonged cooling to allow olivine to reequilibrate at comparatively low temperature. On Earth, lherzolites crystallize either at depths of >8 km (mantle rocks) or as cumulates in large magma chambers. Harvey *et al.* (1993) concluded that the trace element and minor element patterns of LEW88516 and ALH77005 minerals were essentially identical and consistent with large-volume, closed-system fractional crystallization followed by localized crystallization of isolated melt pockets.

2.6.3. *Nakhlites (Clinopyroxenites)*

The cumulate textures of the nakhlites, combined with the presence of lamellar inclusions of augite and magnetite, require slow cooling under highly oxidizing conditions. These textures, especially those of Nakhla and Governador Valadares, are analogous to those of terrestrial augite-rich igneous cumulate rocks of the Abitibi greenstone belt of northern Ontario (Treiman, 1987), where augite cumulates comprise the lower half of a 125 m thick flow. Augite cumulates in the middle third of a 300 m thick sill have little mesostasis, giving them textures more comparable to those of Lafayette. Treiman (1987) concluded from these comparisons that the nakhlites crystallized in thick flows, >125 m thick, or in shallow intrusions, probably less than 1 km deep, of basaltic or picritic magmas. He noted that volcanoes with thick lava flows and evidence of shallow intrusions were common in the Tharsis region of Mars. Furthermore, greenstone belt volcanism may be related to mantle hot spots, another potential analogy to the Tharsis region.

2.6.4. *Chassigny (Dunite)*

The texture and high modal abundance of olivine suggest that Chassigny is a cumulate, also. Floran *et al.* (1978) described the crystallization history as similar to that of nakhlites, except that olivine is much more abundant, and chromite, absent from naklites, crystallizes early. These characteristics suggest that Chassigny and the nakhlites might represent different parts of the same or similar layered igneous complexes. From their REE abundances, Wadhwa and Crozaz (1994) concluded that they could not have crystallized from the same magma, however.

2.6.5. *ALH84001 (Orthopyroxenite)*

Mittlefehldt (1994) interpreted the orthopyroxene and chromite in ALH84001 as cumulus phases. Their textural features indicate slow cooling either during magmatic crystallization, or metamorphic recrystallization, or both. Mineral compositions in ALH84001 are similar to those of lherzolitic shergottites or nakhlites. Mittlefehldt (1994) cited the uniform pyroxene compositions, unusual for Martian meteorites, as indicating that ALH84001 cooled more slowly than did the shergottites, nakhlites, or Chassigny; i.e., it formed at greater depth than they did. Kring and Gleason (1997) argued that the orthopyroxene-silica assemblage present in ALH84001 corresponded to magmatic temperatures of $\sim 1400 - 1470$°C, and to a static pressure of ~0.5 GPa, equivalent to a depth of ~40 km on Mars. Gleason *et al.* (1997) noted that its texture was reminiscent of those of cataclastic anorthosites from the ancient, heavily-cratered, lunar highlands. They, like Treiman (1995b), suggested that at least some of the secondary carbonate formed by replacement of plagioclase, and cite textural evidence as showing this occurred after plagioclase had been converted to maskelynite. Kring *et al.* (1998) concurred in that suggestion, noting that it implied formation of the carbonate after 3.92 ± 0.04 Gyr ago, the time of Ar-degassing of plagioclase according to Turner *et al.* (1997). However, Scott (1999) alternatively suggested that the original carbonates formed as evaporite deposits, probably prior to impact heating ~4 Gyr ago, when episodic floods were more common. He suggests that preservation of the carbonates for ~4 Gyr was aided by the impact, which sealed up the carbonate-bearing fractures and pores, making the rock less pervious to later infiltration of fluids.

3. Shock Metamorphism

3.1. Evidence of Shock

It is generally agreed that the Martian meteorites have been ejected from the planet's surface by large-scale impacts. The ejection velocity must have exceeded the escape velocity of Mars, which is about 5 km/s. Material accelerated by a shock wave to >5 km/s should be in a molten state according to basic shock wave physics. The fact that all known Martian meteorites are solid though strongly shocked rock fragments prompted Melosh (1984) to develop a model of the ejection process in which a special spallation mechanism provides most of the required ejection velocity for rock fragments ejected from a thin, uppermost layer of the impacted target without melting them. Although alternative mechanisms have been proposed (Nyquist, 1983; O'Keefe and Ahrens, 1986), this model has been widely accepted. The originally proposed spallation mechanism (Melosh, 1984) required the parent craters of the meteorites to be larger than ~10 km in diameter. Recent refinements of the model have reduced the size limit to >3 km (Head and Melosh, 2000).

As expected from the impact and ejection model for the origin of the Martian meteorites, the imposed extreme physical conditions caused significant changes in the textures, mineralogy, and possibly even the isotopic compositions of constituent mineral phases. It even led to a shock- induced implantation of Martian atmospheric gases into the meteorites (Duke, 1968; Stöffler *et al.*, 1986; Bogard *et al.*, 1986; Wiens and Pepin, 1988; McSween, 1994). *The important obervation is that all Martian meteorites are moderately to strongly shock metamorphosed by shock pressures ranging between about 15 and 45 GPa*. The understanding of the type and intensity of shock metamorphism of Martian meteorites is thus essential for the interpretation of the ejection and possible impact-induced relocation processes, which relate to some extent to the problem of the geological provenance, and to the interpretation of analyzed isotope systems, which may be disturbed by shock.

It has been recognized since the pioneering studies of Tschermak (1872) that shergottites and some other achondrites are severely shocked (Binns, 1967; Duke, 1968). Although shock effects in meteorites were known before the Martian origin of the SNC meteorites was suspected (e.g., Wood and Ashwal, 1981), the Martian origin hypothesis gave new impetus to their study. The degree of shock metamorphism in shergottites was first studied quantitatively on Shergotty (Lambert and Grieve, 1984; Stöffler *et al.*, 1986). In all shergottites the constituent minerals display specific, more or less similar shock effects, well known from naturally and experimentally shocked terrestrial, lunar, and meteoritic rocks (Stöffler, 1972; Stöffler *et al.*, 1988; Bischoff and Stöffler, 1992). Pyroxene shows strong mosaicism, mechanical twinning, shear fractures and various lattice defects such as high dislocation densities revealed in the Transmission Electron Microscope (TEM; e.g., Müller, 1993). Olivine, if present, is affected by strong mosaicism, deformation bands, planar fractures, planar deformation features and high dislocation densities (e.g., Greshake and Stöffler, 1999, 2000). Ostertag *et al.* (1984) attributed the brown staining of olivine in ALH77005 to a shock-induced oxidation of iron to Fe^{3+}. Plagioclase is transformed to diaplectic glass (maskelynite) and retains its primary crystal shape. Based on the experimentally calibrated refractive index of maskelynite, the peak shock pressure (final equilibrium shock pressure) of the host meteorite can be deduced (Stöffler *et al.*, 1986).

A typical feature of the shergottites is the presence of shock-produced veins and melt pockets caused by local pressure and temperature excursions of presumably up to 60 – 80 GPa and 2000 °C (Stöffler *et al.*, 1986). These pressure estimates are based on experimental data (Kieffer *et al.*, 1976; Schaal and Hörz, 1977; Schmitt, 2000). In addition, high pressure phases such as very dense post-stishovite polymorphs of SiO_2 have been discovered in Shergotty (Sharp *et al.*, 1999; El Goresy *et al.*, 2000). It has to be pointed out that these high pressures were only produced very locally and do not represent the equilibration pressure as previously suggested (Sharp *et al.*, 1999). Recently, an assemblage of omphacite, stishovite and $KAlSi_3O_8$-hollandite was found in a shock vein of Zagami, indicating crystallization of these phases during decompression between 25 and 50 GPa (Langenhorst and Poirier, 2000). Also, the high-pressure polymorphs of olivine and pyroxene, ringwoodite and majorite, were tentatively reported from a melt vein in the basaltic shergottite EETA79001 (Steele and Smith, 1982). However, unambiguous identification of these two phases has failed so far (Boctor *et al.*, 1998). Moreover, the melt pockets in some Martian meteorites are obviously the host regions of gases of the Martian atmosphere that were first detected in EETA79001 (Bogard and Johnson, 1983; Becker and Pepin, 1984). These gases must have been implanted during shock metamorphism of the meteorite precursor rocks near the Martian surface. Experimental studies on shock implantation show that shock can relatively easily incorporate an ambient gas phase into solid material, even at temperatures well below melting (Bogard *et al.*, 1986; Wiens and Pepin, 1988).

Refractive index measurements of maskelynite gave quantitative estimates of the peak shock pressure for some shergottites: Shergotty: 29 ± 1 GPa (Stöffler *et al.*, 1986), Zagami: 31 ± 2 GPa (Stöffler *et al.*, 1986; Langenhorst *et al.*, 1991), EET79001: 34 ± 2 GPa (Lambert, 1985), ALH77005: 43 ± 2 GPa (McSween and Stöffler, 1980). For other shergottites the values, based on the overall shock effects in plagioclase, olivine, and pyroxene, and on the presence and abundance of localized melts, are less accurate: Dar al Gani 476: probably $30-35$ GPa (Greshake and Stöffler, 1999, 2000), QUE94201: $30-35$ GPa, Los Angeles: $35-40$ GPa, Dhofar 019: $35-40$ GPa, Sayh al Uhaymir: $35-40$ GPa, LEW88516: ca. $40-45$ GPa ("strongly shocked", Keller *et al.*, 1992), Y793605: ca. $40-45$ GPa. All estimated peak shock pressures of shergottites are summarized in Table I, along with the estimated post-shock temperatures.

Considering the range of shock pressures observed in shergottites, it is conspicuous that the basalts were all affected by similar shock pressures in the range of $\sim 30-35$ GPa, whereas the lherzolites reveal somewhat higher shock pressure ($\sim 40-45$ GPa). Also, Ott and Löhr (1992) noted that the ^{4}He content of lherzolite LEW88516 indicates complete loss of radiogenic ^{4}He acquired prior to its ejection from Mars ~ 3 Myr ago, consistent with its high post-shock temperature of $\sim 600\,^\circ$C (Table I). The type and homogeneity of shock damage observed in the constituent minerals of the basaltic shergottites indicates that each of them was affected by only one impact event (Stöffler *et al.*, 1986; Müller, 1993). This seems to be different for the lherzolitic shergottites, as observed for Y793605, which has been brecciated by a first impact and shock metamorphosed by a second impact. The three nakhlites are less intensely affected by shock metamorphism than the other Martian meteorites. Only weak undulatory extinction and a rather low dislocation density in olivine as well as entirely birefringent plagioclase suggest a peak shock pressure of $\sim \leq 20 \pm 5$ GPa (Bunch and Reid, 1975; Greshake, 1998).

Shock metamorphism in Chassigny was investigated in detail by optical and transmission electron microscope (Langenhorst and Greshake, 1999). Conversion of feldspars to diaplectic glass (maskelynite), the clino-/orthoenstatite inversion, strong mosaicism of olivine, and the activation of numerous planar fractures and c-dislocations in olivine are among the shock effects. High-resolution TEM revealed additionally the coexistence of planar fractures with discontinuous fractures in olivine. These findings point to a shock pressure of about 35 GPa.

TABLE I

Estimates of the peak shock pressure (final equilibration shock pressure) and the overall post-shock temperature increase in Martian meteorites. Data from Stöffler *et al.* (1986) and Stöffler (2000) except for Sayh al Uhaymir 005, Los Angeles, and Dhofar (this paper).

Meteorite	Shock pressure (GPa)	Post-shock temperature*
Shergotty	29 ± 1	200 ± 20
Zagami	31 ± 2	220 ± 50
EETA 79001	34 ± 2	250 ± 50
QUE94201	~30–35	~200 – 350
Dar al Gani 467	~35–40	~350 – 450
Los Angeles	~35–40	~350 – 450
Dhofar 019	~35–40	~350 – 450
Sayh al Uhaymir 005	~35–40	~350 – 450
ALHA77005	43 ± 2	~450 – 600
LEW88516	~45	~600
Y793605	~45	~600
ALH84001	~35–40	~300 – 400
Nakhlites	~20 (±5)	~100
Chassigny	~35	~300

*Relative to ambient pre-shock temperature.

In the orthopyroxenite ALH84001, shock metamorphism is documented by complex textures, such as localized brecciation in fine-grained shear zones, strong mosaicism and numerous irregular fractures in orthopyroxene, and by the conversion of all plagioclase to maskelynite. While Mittlefehldt (1994) explained these effects by a single impact event, Treiman (1998) invoked up to five impacts. We believe that the presence of maskelynite in both brecciated and non-brecciated regions indicates that at least two impact events are required: A first weak shock event producing the brecciation and a subsequent stronger shock event which transformed plagioclase to maskelynite throughout the whole rock.

3.2. Environments and Implications of Shock Metamorphism

The observed shock metamorphism of Martian meteorites has important implications for their impact and ejection history and for their geologic provenance, if the geologic settings of their magmatic formation processes are taken into account. Summarizing the essential observations leads us to some general conclusions.

All Martian meteorites are moderately to severely shocked (Table I, Figure 7), with effects being homogeneously distributed throughout the rocks. These shock effects can be attributed to one specific event in each case, most probably the ejection event. A single stage shock history is implied for all basaltic shergottites and most likely for the nakhlites and for Chassigny. However, some of the ultramafic "plutonic" rocks such as lherzolitic shergottite Y-793605 and orthopyroxenite ALH84001 are shocked monomict breccias indicating a two-stage shock history: In a first impact, the rocks are brecciated at very low shock pressure at depth and relocated to the surface during the same event as commonly observed in terrestrial impact craters such as the Ries (e.g., Pohl *et al.*, 1977). The transformation of plagioclase to maskelynite indicates strong shock metamorphism in a second impact, most probably the ejection event. Although clear evidence for the two remaining lherzolitic basalts and for the nakhlites/chassigny group is unavailable, an impact-induced relocation of the"plutonic" ultramafic Martian meteorites from their primary deep-seated magmatic setting is highly plausible.

The second fundamental observation relates to the ranges of observed shock pressures for all the Martian meteorites and for particular groups of them (Table I, Figure 7). Although exact values are not yet available, we recognize 1) that the observed shock pressures are restricted to a range of about 15 to 45 GPa, and 2) that the basaltic shergottites range from about 30 to 35 GPa, the lherzolitic shergottites from about 40 to 45 GPa, and the nakhlites from about 15 to 25 GPa. This means that unshocked meteorites as well as shock-fused meteorites are lacking and that the observed 15–45 GPa range may be viewed as a typical "launch window" for Mars. The lower limit may indicate that unshocked rocks and rocks shocked to pressures lower than the Hugoniot Elastic Limit cannot be ejected, in contrast to what has been proposed by Melosh (1995), Mileikowsky *et al.* (2000), and Weiss *et al.* (2000). The upper limit indicates that melt ejecta are too much dispersed and, hence, too small to survive as meteoroids. Comparing meteorites from Mars, the Moon, and the eucrite parent body, it seems that the observed range of shock metamorphism related to the ejection event is a function of the size of the parent planetary body and therefore of the magnitude of the escape velocity: The present data indicate that lunar meteorites are shocked below about 20 GPa (Bischoff and Stöffler, 1992; Greshake *et al.*, 2001), and meteorites of the eucrite-howardite-diogenite group, possibly originating from the 550 km diameter asteroid Vesta, are at most mildly shocked, i.e. below ∼5–10 GPa (Metzler *et al.*, 1995).

A third implication of the observed shock metamorphism of Martian meteorites relates to the size of the precursor meteoroids and to their ejection ages. As known from terrestrial craters such as the Ries crater (Pohl *et al.*, 1977; Stöffler and Ostertag, 1983; von Engelhardt and Graup, 1984), the size of displaced shocked rock fragments is inversely proportional to the shock intensity. Crystalline rock fragments in polymict breccias of the Ries shocked to the range of the Martian shergottites (∼30−45 GPa) do not exceed 0.5 m or so, and most sizes are ∼0.1−10 cm. Typical shock stage III rocks (45−60 GPa; Stöffler, 1984) are consistently <50 cm in size. Additionally, distal ejecta (solid clasts) in the Ries (Reutter blocks: Upper Jurassic limestone fragments, Pohl *et al.*, 1977) are not only small, <∼20 cm, but are also derived from the uppermost layer of the target in agreement with the conditions invoked by the spallation model (Melosh, 1984). Consequently, the lherzolitic shergottites (40−45 GPa) must have originated from ≈0.1 m-sized rocks and cannot have been ejected in one block together with those basaltic shergottites that have the same ejection age (Figure 7a). Rather, the basalts and lherzolites may be derived from different surface regions of the same parent crater notwithstanding the fact that the lherzolites had to be relocated first to the surface by a previous impact. A similar case could be made for the nakhlites and Chassigny, which also have identical ejection and crystallization ages but different shock pressures.

The peak shock pressures experienced by the Martian meteorites are likely to be a consequence of their geometrical relationship to "ground zero" at the moment of the impact that is destined to launch them from the planet. The near-surface spall model, for example, outlines rather definite relationships between the impactor, the transient crater cavity, and the target region near ground zero (Melosh, 1984; Figure 11). Those fragments destined for ejection might be considered to constitute the "lid" of the transient cavity; a lid destined to be blown off. In the spallation model, the thickness of the "lid" is given by the depth of the spall zone, and was estimated by Warren (1994) to be 0.2−0.4 times the diameter of the projectile at a distance of 1−3 projectile radii from the the impact. Thus, the "lid" would be on the order of 50 to 100 m thick for a 10 km diameter crater. (See "Potential source terrains" later in the paper). In the lid, peak shock pressure increases in the downward direction from the surface, and decreases in the radial direction from ground zero. The nakhlites, being most lightly shocked, are thus implied to have been ejected from nearest the Martian surface, in spite of having probably crystallized near the center of a thick flow, $>\sim$100 m thick, or in a subsurface intrusion. Chassigny, being more severely shocked, is implied to have been ejected from a deeper region of the lid, if the nakhlites and Chassigny were ejected simultaneously. The lherzolitic shergottites are most highly shocked of all the Martian meteorites, and thus are expected to come from deep within the lid, close to the melt zone. If, for example, they and the basaltic shergottites were ejected simultaneously, the latter would have come from nearer to the surface, consistent with being recent lava flows. The orthopyroxenite ALH84001, which likely crystallized at the greatest depth of the Martian meteorites, experienced peak shock pressure equivalent to those of the shergottites, implying prior excavation from depth to the launch site. This is consistent with an ancient age and origin in the Martian highlands, which probably were "gardened" to depths on the order of a kilometer, or more (W. Hartmann, personal communication). Gardening of the surfaces of ∼180 Myr old basaltic shergottite lava should have been minimal, however, and shergottites are likely to have been ejected from their place of emplacement as lava flows.

Finally, we note that rocks of distinctly different shock pressures, e.g. nakhlites and Chassigny, or basaltic and lherzolitic shergottites, cannot have been ejected from Mars in one large rock unit. Such scenarios have been proposed in order to explain the different exposure ages within the shergottite group as due to later break-up in space. Indeed, the limited size of strongly shocked rocks ejected from

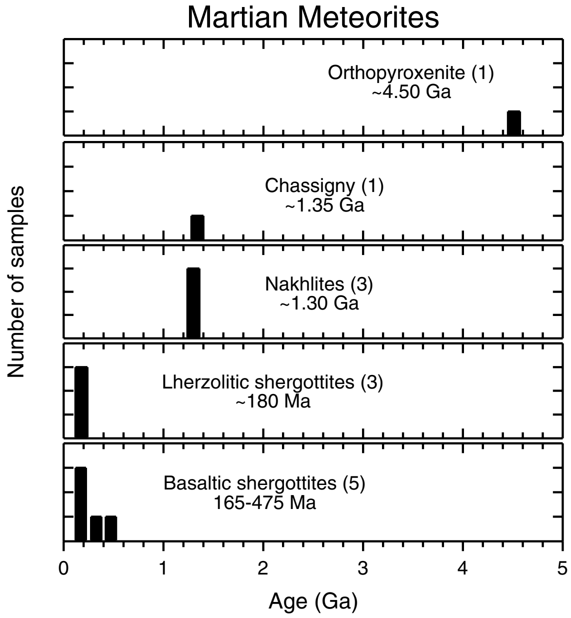

Figure 2. The crystallization ages of Martian meteorites separated by compositional group. The values plotted are the "preferred ages" from Tables II and III. The oldest meteorite in the Martian clan is ~4.5 Gyr old, and the youngest ~180 Myr old. Thus, Martian magmatism appears to have extended over most of solar system history, a conclusion that agrees with the time span of crater retention ages (Hartmann and Berman, 2000). The thirteen meteorites fall into only five age groups, leaving large gaps in Martian chronology as recorded by the meteorites.

the parent crater (see above) also argues against this possibility. In the case of the basaltic shergottites, however, the peak shock pressures are nearly equivalent, leaving the comparatively large size required of the initial ejecta fragments as the primary physical limitation on secondary break-up scenarios. In later sections, we will further discuss the issue of the number of required ejection events, in conjunction with the problem of the geological provenance of the Martian meteorites.

4. Radiometric Ages

Radiometric ages of Martian meteorites as reported in the literature are given in Tables II and III. We discuss the ages of individual meteorites within each of the meteorite classes separately. For the nakhlites and Chassigny (Table II), the ages determined by the various radiometric methods are in close agreement and present a coherent picture of when those rocks crystallized as thick magma flows or subsurface sills. For the other classes of Martian meteorites, notably the shergottites (Table III), the picture is more complicated.

Figure 2 gives an overview of those data that most reliably give the crystallization ages of the meteorites. The 13 meteorites define only 5 separate ages, covering

TABLE II
Summary of Radiometric Ages of Martian Meteorites: Nakhlites, Chassigny, ALH84001

Meteorite	K-Ar (Gyr)	^{39}Ar-^{40}Ar (Gyr)	Rb-Sr (Gyr)	Sm-Nd (Gyr)	U-Th-Pb (Gyr)	Preferred Age (Gyr)
Clinopyroxenites (Nakhlites):						
Nakhla	1.30 ± 0.03[a]	1.3[b]	1.23 ± 0.01[c] 1.30 ± 0.02[e] 1.36 ± 0.02[e]	1.26 ± 0.07[d]	1.28 ± 0.05[d] 1.24 ± 0.11[d]	**1.27 ± 0.01**
Governador Valadares		1.32 ± 0.04[f]	1.32 ± 0.01[g] 1.19 ± 0.02[h]	1.37 ± 0.02[h]		**1.33 ± 0.01**
Lafayette		1.33 ± 0.03[b]	1.25 ± 0.08[i]	1.32 ± 0.05[I]		**1.32 ± 0.02**
Dunite:						
Chassigny	1.39 ± 0.17[j]	1.32 ± 0.07[k]	1.22 ± 0.01[l]	1.36 ± 0.06[m]		**1.34 ± 0.05**
Orthopyroxenite :						
ALH 84001						
Silicates		3.92 ± 0.10[n] 4.07 ± 0.04[o] 4.10 ± 0.20[p]	4.55 ± 0.30[q] 3.89 ± 0.05[r]	∼4.56[s] 4.50 ± 0.12[q]		**4.51 ± 0.11**
Carbonates		∼3.6[t]	3.90 ± 0.04[u] 1.41 ± 0.10[r]		4.04 ± 0.10[u]	**3.92 ± 0.04**

References: [a]Stauffer (1962); [b]Podosek (1973); [c]Gale *et al.* (1975); [d]Nakamura *et al.* (1982a); [e]Papanastassiou and Wasserburg (1974); [f]Bogard and Husain (1977); [g]Wooden *et al.* (1979); [h]Shih *et al.* (1999); [i]Shih *et al.* (1998); [j]Lancet and Lancet (1971); [k]Bogard and Garrison (1999); [l]Nakamura *et al.* (1982b); [m]Jagoutz (1996); [n]Turner *et al.* (1997); [o]Ilg *et al.* (1997); [p]Bogard and Garrison (1999); [q]Nyquist *et al.* (1995); [r]Wadhwa and Lugmair (1996); [s]Jagoutz *et al.* (1994); [t]Knott *et al.* (1996); [u]Borg *et al.* (1999).

an age span from the formation of the planet extending nearly to the present day, and there is only a single Martian rock older than 1.3 Gyr. This is a rather incomplete sample of the ages of Martian surface rocks, and these ages only give a record of Martian evolution, if their geologic context is known. Isotopic data of Martian meteorites are most useful to study the Martian global geochemical evolution.

Radiometric ages, though, provide absolute calibration marks for relative ages from cratering records. By extrapolating the lunar cratering rate to Mars, the relative ages of Martian surface units can be estimated from the density of craters on them (Neukum *et al.*, 2001; Ivanov, 2001; Hartmann and Neukum, 2001). These ages are divided into three major chronostratigraphic units, or epochs: Noachian, Hesperian, and Amazonian. The Noachian and Amazonian are further subdivided into Early, Middle, and Late periods; whereas the Hesperian is simply divided into Early and Late periods (Tanaka, 1986; Tanaka *et al.*, 1992). We begin our discussion with early Mars, working towards the present day.

TABLE III

Summary of Radiometric Ages of Martian Meteorites: Shergottites

Meteorite	K-Ar (Myr)	^{39}Ar-^{40}Ar (Myr)	Rb-Sr (Myr)	Sm-Nd (Myr)	U-Th-Pb (Myr)	Preferred Age (Myr)
Shergottittes (Basalts) :						
Shergotty	580 ± 50[a1] 196 ± 40[w]	254 ± 10[b] 167[c]	163 ± 12[d] 165 ± 4[e]	147 ± 20[e] 360 ± 16[e] 620 ± 171[f]	200 ± 4[g] 437 ± 36[g] 600 ± 20[g] 217 ± 110[h] 189 ± 83[h]	**165 ± 4**
Zagami		242[c]	178 ± 3[f] 174 ± 14**i** 163 ± 19[i]	163 ± 7[i]	230 ± 5[g] 229 ± 8[g]	**177 ± 3**
Los Angeles			165 ± 11[j]	172 ± 8[j]		**170 ± 8**
EETA79001A		2035[c]	172 ± 18[k]		150 ± 15[g] 170 ± 36[g]	**173 ± 3**
EETA79001B			177 ± 12[k] 173 ± 3[l]	165 ± 43[l]		
QUE94201		730[c]	327 ± 12[m]	327 ± 19[m]		**327 ± 10**
DaG476				474 ± 11[n] ~800[o]		**474 ± 11**
Shergottites (Lherzolites) : :						
ALHA77005	1330 ± 130[p]	3500[c]	156 ± 6[q] 188 ± 11[f] 185 ± 11[r]	173 ± 7[r]		**179 ± 5**
LEW88516	2600[c]		183 ± 10[s]	166 ± 16[t]	~170[u]	**178 ± 8**
Y793605		1595[c]			212 ± 62[v]	**212 ± 62**

References: [a1]Geiss and Hess (1958), recalculated to the K-decay constants by Steiger and Jäger (1977); [a]Eugster *et al.* (1997a); [b]Bogard *et al.* (1979); [c]Bogard and Garrison (1999); [d]Nyquist *et al.* (1979a); [e]Jagoutz and Wänke (1986); [f]Shih *et al.* (1982); [g]Chen and Wasserburg (1986); [h]Sano *et al.* (2000); [i]Nyquist *et al.* (1995); [j]Nyquist *et al.* (2000); [k]Nyquist *et al.* (1986); [l]Nyquist *et al.* (2001); [m]Borg *et al.* (1997); [n]Borg *et al.* (2000); [o]Jagoutz *et al.* (1999), Jagoutz and Jotter (2000); [p]Miura *et al.* (1995); [q]Jagoutz (1989); [r]Borg *et al.* (2001b); [s,t]Borg *et al.* (1998a, 1998b); [u]Chen and Wasserburg (1993); [v]Misawa *et al.* (1997); [w]Terribilini *et al.* (1998).

4.1. Orthpyroxenite ALH84001: A Carbonate-Bearing Fragment of the Noachian Crust

ALH84001 is the only meteorite from the ancient Martian crust. We infer its crustal origin from its ancient age, and less directly from its composition. The old crystallization age of ALH84001 is direct evidence that portions of the Martian crust formed quickly after the planet accreted. There have been some variations in the ages reported for ALH84001 (Figure 3). Jagoutz *et al.* (1994) argued that the Sm-

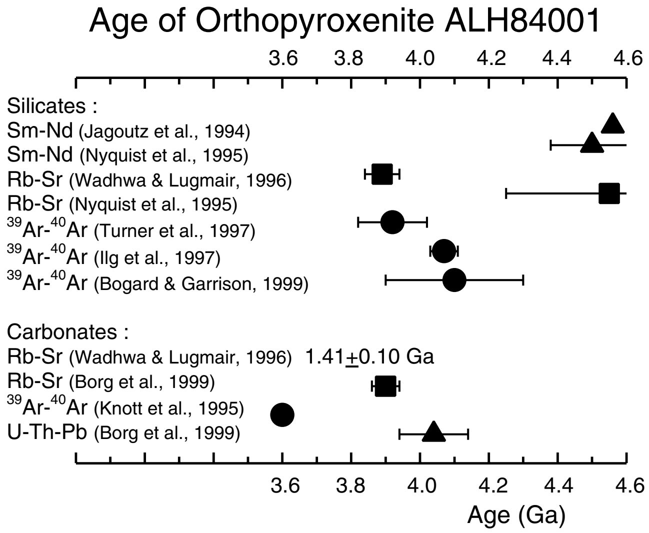

Figure 3. Radiometric ages of orthopyroxenite ALH84001, as determined by a variety of techniques. Sm-Nd isochron ages for silicate minerals indicate primary crystallization prior to 4.4 Gyr ago. ^{39}Ar-^{40}Ar ages for the silicates indicate Ar-outgassing between ~3.9–4.1 Gyr ago in a secondary heating event, presumably related to impact cratering. Rb-Sr ages of different subsamples have given different results, apparently related to both primary crystallization and secondary reheating. Attempts to date secondary carbonates within ALH84001 have yielded ages as low as ~1.4 Gyr and as high as ~4.0 Gyr. We prefer a carbonate age of ~3.9 Gyr, which would make carbonate formation directly or indirectly related to the cratering event that reset the ^{39}Ar-^{40}Ar age.

Nd isotopic data plotted along a 4.56 Gyr reference isochron. However, they neither reported an isochron regression nor estimated an uncertainty on the age. Their data were for bulk samples and acid leach/residue pairs, and thus any “isochron” also may be interpreted as an “unmixing” line. This is because phosphate minerals, major contributors to the REE budget, would be dissolved into the “leach” solutions, leaving the residue as a complementary end member. Nyquist *et al.* (1995) reported both Sm-Nd and Rb-Sr data for ALH84001. They obtained a ^{147}Sm-^{144}Nd isochron age of 4.50 $\pm$ 0.13 Gyr from a suite of bulk samples and a pyroxene mineral separate. Their isochron also could be interpreted as a mixing line between orthopyroxene and a second component of low Sm-Nd ratio, like phosphate, if that component dominated the REE budget of the rock. However, both phosphate and plagioclase have low Sm/Nd ratios, and contribute to the REE budget. Because both are in low abundance, both could be randomly distributed among the different bulk samples analysed, decreasing the likelihood that the isochron is simply a two-component mixing line. Nyquist *et al.* (1995) also determined initial (^{146}Sm/^{144}Sm)$_I$ = 0.0022 $\pm$ 0.0010 for ALH84001 from variations in ^{142}Nd/^{144}Nd caused by decay of ^{146}Sm ($T_{1/2}$ = 103 Myr), initially present in the rock, to ^{142}Nd. This result suggests that ALH84001 formed more than one half-life of ^{146}Sm after the angrite meteorite LEW86010, for which (^{146}Sm/^{144}Sm)$_I$ was ~0.0070–

0.0076 (Lugmair and Galer, 1992; Nyquist *et al.*, 1994). The ^{146}Sm/^{144}Sm ratio requires closure of the Sm-Nd system in ALH84001 no earlier than ~115 Myr after formation of the angrite. Although ^{142}Nd/^{144}Nd measurements are analytically challenging, the long- and short-lived chronometers can be considered concordant for an age of ~4.4 Gyr, the lower limit on the conventional ^{147}Sm-^{144}Nd age. Thus, the great antiquity of ALH84001 appears to be established, in spite of generally lower ^{39}Ar-^{40}Ar ages, and a lower Rb-Sr age (Wadhwa and Lugmair, 1996).

The ^{39}Ar-^{40}Ar ages of ALH84001 are in the range ~3.8–4.2 Gyr (Figure 3). Thus, Ar-outgassing appears to have occurred after the parental rock solidified. Turner *et al.* (1997) derived an ^{39}Ar-^{40}Ar age of 3.92 ± 0.10 Gyr, whereas Ilg *et al.* (1997) reported an older age of 4.07 ± 0.04 Gyr. Martian Ar components in ALH84001 are difficult to characterize, and correcting for the uncertainty in trapped Martian ^{40}Ar allows ages in the broader interval 4.10 ± 0.20 Gyr (Bogard and Garrison, 1999). The time when ALH84001 was outgassed corresponds to the hypothesized period of the "terminal cataclysm" on the moon and HED parent body (Bogard, 1995). Perhaps Mars also experienced such a cataclysmic bombardment, but this remains a tentative conclusion, based on a single sample.

The Rb-Sr age of 3.84 ± 0.05 Gyr for ALH84001 (Wadhwa and Lugmair, 1996) is significantly younger than the Sm-Nd age and also than the Rb-Sr age of Nyquist *et al.* (1995). Apparently, the secondary reheating event that reset the ^{39}Ar-^{40}Ar age affected different portions of the rock to different degrees. The rock contains crushed zones (Treiman, 1995b) that may have been more severely affected by the impact event than were intact orthopyroxenite areas. Furthermore, secondary carbonate mineralization is preferentially found within these crushed zones. Thus, we concur with the interpretation of Wadhwa and Lugmair (1996) that the young Rb-Sr age of ~3.9 Gyr (^{87}Sr decay constant $\lambda_{87} = 1.402 \times 10^{-11}$ yr^{-1}, Minster *et al.*, 1982) represents a time of intense shock and post-shock thermal annealing. This interpretation implies that the Rb-Sr ages of some portions of the rock also were reset by the impact of a large meteoroid on Mars. Both interpretations may apply, as ALH84001 bears evidence of several major meteoroid impacts (*cf.* Treiman, 1998), not surprising for a rock from the Martian highlands, which have been "gardened" by meteoroid impact to a depth of ~1 km (Hartmann *et al.*, 2000).

ALH84001 contains ~1 vol.% of secondary carbonates. The secondary Sr or Nd in these carbonates may have disturbed the isotopic systems. The work of Borg *et al.* (1999), discussed below, shows that the leaching procedure used by Jagoutz *et al.* (1994) to determine the Sm-Nd age of ALH84001 would have dissolved both igneous phosphates and secondary carbonates. Also, the isochron of Nyquist *et al.* (1995), determined by bulk samples plus orthopyroxene, would be subject to variations in the relative proportions of phosphates and carbonates. The REE abundances in the primary phosphates probably are much higher than in the secondary carbonates, however, and it is likely that the presence of carbonates has not significantly affected the Sm-Nd isochrons.

Three attempts to date the carbonates have been reported (Figure 3). Knott *et al.* (1996) reported an age of ~3.6 Gyr by laser-probe ^{39}Ar-^{40}Ar dating. Turner *et al.* (1997), however, interpreted those data as heavily influenced by outgassing from the plagioclase substrate beneath the carbonate grain they analysed. Wadhwa and Lugmair (1996), adopting the model of Treiman (1995b) for formation of the carbonates by replacement of plagioclase, proposed an age of 1.41 ± 0.10 Gyr for the carbonates from a two-point carbonate-plagioclase "isochron". However, the Sr-isotopic composition of plagioclase is variable, making pairing of carbonate and plagioclase for dating ambiguous. In the third investigation of the carbonate age, Borg *et al.* (1999) exploited the compositional zoning of the carbonate minerals in ALH84001 to selectively dissolve phases having different parent/daughter ratios for Rb-Sr, U-Pb, and Sm-Nd dating. Although REE concentrations in the resultant solutions were too low for Nd isotopic analysis, the Rb-Sr and U-Pb isotopic analyses yielded concordant ages of ~3.9–4.0 Gyr, close to those originally obtained by laser probe ^{39}Ar-^{40}Ar dating (Knott *et al.*, 1996). These results, combined with the ^{39}Ar-^{40}Ar studies of ALH84001 silicates, suggest that plagioclase outgassing and carbonate formation were contemporaneous, and possibly even simultaneous. If so, the lower ^{39}Ar-^{40}Ar age reported for "carbonate" by Knott *et al.* (1996) may reflect some ^{40}Ar loss from this low-temperature secondary mineral phase.

Differences in interpretation of the radiometric age data for the carbonates may be related to the fact that there are several types, and possibly two or more generations, of carbonates present in ALH84001. Mittlefehldt (1994) identified two generations, "early" (pre-shock) carbonates, and "late" (post-shock) carbonates. Treiman (1995b) suggested the carbonates formed via replacement of plagioclase, a suggestion that strongly influenced the Rb-Sr study of Wadhwa and Lugmair (1996), as well as interpretation of the ^{39}Ar-^{40}Ar study of Knott *et al.* (1996). Gleason *et al.* (1997) and Kring *et al.* (1998) also favored carbonate formation via dissolution-replacement reactions between CO_2-charged fluids and maskelynite. They present as evidence carbonates filling small pockets in pyroxene previously occupied by maskelynite, as seen in photomicrographs of a thin section of the meteorite (Gleason *et al.*, 1997; Figure 4). Kring *et al.* (1998) argue from an electron microprobe study of K and Ca in six different complexly zoned carbonate patches in a single thin section that the laser probe ^{39}Ar-^{40}Ar study of Knott *et al.* (1996), as reported by Turner *et al.* (1997), does not give the age of the carbonates. They reached this conclusion because most of the data for which carbonate was identified as the target in the study of Turner *et al.* (1997) showed the presence of more K than could be accounted for by carbonates alone in the electron probe study. However, carbonates ~100 μm in diameter also are easily visible with a binocular microscope along fractured surfaces of macroscopic pieces of the meteorite. There usually is no visible association with comparatively rare maskelynite, although sometimes such an association does exist. (The proportions of maskelynite and carbonate are subequal at ~1%). These latter carbonates are of the globular variety, a photomicrograph of which was shown by McKay *et al.* (1996). These fracture-filling carbonates are described very completely in the paper by Scott *et al.* (1998), and were the intended objects of the investigation by Borg *et al.* (1999) of carbonate fragments picked from ∼1 g of the meteorite.

It seems probable that some of the confusion concerning interpretation of the carbonate ages stems from occasionally inappropriate application of observations made on limited samples of the meteorite. Here, we follow most closely the discussion of Scott *et al.* (1998), who examined nine polished thin sections of the meteorite. Quoting: "Carbonates in ALH84001 occur in three distinct locations: in pyroxene fractures, in crushed zones (also called granular bands; Treiman, 1995b), and as massive grains and globules on pyroxene grain boundaries (e.g., Mittlefehldt, 1994; Treiman, 1995b) .... Carbonates in pyroxene fractures can be divided conveniently into three types according to their shape and nature of the fractures in which they formed: disks, dike-shaped veins, and irregularly shaped grains." These authors (and others) document that all types of carbonates are similarly compositionally zoned. The carbonates nucleated with Ca-rich cores and became richer in Fe and then Mg as they grew outward. Last to form were the magnesite rims. It is this zonation that the experiment of Borg *et al.* (1999) was designed to exploit: Enrichment of Sr and Pb over Rb and U in the Ca-rich cores, leaving enhanced Rb/Sr and U/Pb ratios in the last-formed magnesites. Similar zoning profiles in all types of carbonates imply that carbonate formation took place as a single event. Again quoting Scott *et al.*: "...there is much evidence that carbonates in fractures did not form by replacement of plagioclase glass."

The simplest interpretation of the observations appears to be:

1. Some carbonates formed by replacement reactions with crystalline plagioclase. A probable example is seen in Figure 4d of Gleason *et al.* (1997). If, as argued by the authors and Kring *et al.* (1998), replacement was of maskelynite, a prior shock event is required.
2. Not all carbonates formed by replacement reactions (Scott *et al.*, 1998). The majority of carbonates probably formed without need of plagioclase or maskelynite, but if some were present, reactions could occur. CO_2-enriched aqueous fluids apparently circulated through the rock, implying the prior existence of a fracture network. Thus, the rock had been brecciated prior to that time, probably by excavation from great depth to a surface or near-surface location. The compositional zoning of the carbonates was established at that time, and dated at ∼4.0 Gyr ago by Borg *et al.* (1999).
3. A second shock fractured some carbonates and formed maskelynite and plagioclase glass, some of which can now be found in fractures. This shock also opened up some pre- existing fractures in which carbonates already had formed (Scott *et al.*, 1998, Figure 3a), and resealed others. This last shock is most likely the ejection event. The other Martian meteorites invariably show shock levels of 15–45 GPa, implying that such shock levels are required for their ejection from the planet. Thus, this second shock happened ∼15 Myr ago, the ejection age of ALH84001, as discussed in a later section.

Treiman (1998) suggested a more complex scenario involving 4 "compositional", 6 to 8 "deformational", and 4 "impact" events. A critical difference to the scenario above is that Treiman's last impact event occurs without major shock metamorphism. Attempts to refine the inferred history of ALH84001, including the radiometric age of the carbonates, are likely to continue, but the concordant Rb-Sr and U-Pb ages of Borg *et al.* (1999) seem presently to be preferred.

The mechanism of carbonate formation has been debated in the context of "impact metasomatism" (Harvey and McSween, 1996) and "playa lake" models. Playa lake models have gained favor, seeming to be more consistent with various types of data (*cf.*, Warren, 1998). Also, in an experimental study, Golden *et al.* (2000a, 2000b) were able to reproduce the carbonate zonation profiles in ALH84001. They used a multi-step, sequential aqueous precipitation from fluids of changing composition, followed by a final reheating to 470°C. Whether this process mimics what might happen on Mars, perhaps in a Martian playa lake, has not been addressed in detail. However, one can easily envision a scenario in which Ar-outgassing from ALH84001 accompanied a crater-forming event that left ALH84001 either as part of the crater ejecta blanket, or, in the playa lake model, at the bottom of a crater. Scott *et al.* (1998) have suggested that ALH84001 may have been located beneath the central region of a large impact crater or basin that formed $\sim$4 Gyr ago. The playa lake model would definitely be favored over impact metasomatism if Ar outgassing were earlier than carbonate formation, occuring at most 4.1 Gyr ago. The carbonate age data would then be consistent with precipitation from a crater lake filled slightly later by surface runoff into the crater. Alternatively, formation of a crater lake might be triggered by formation of the crater itself, filled by melted groundwater released by the heat of the impact. Newsom *et al.* (1996) have argued that formation of large ($>$ 65 km diameter) impact craters on Mars may have been accompanied by the creation of ice-covered impact crater lakes, which would not freeze totally over a lifetime of $\sim 10^4$ years. Supply of water to them from deep aquifirs might provide a connection to possible life residing in the aquifirs (Boston *et al.*, 1992). Thus, although the suggestion by McKay *et al.* (1996) that certain worm-like morphological features in ALH84001 might be relics of Martian life has proven controversial, that interpretation is consistent with the apparent age of the carbonates, the time of formation of equivalent lifeforms on earth, and possible access to potential subsurface habitats via crater formation. The crater lake scenario remains speculative, but is consistent with suggested modes of carbonate formation, and is made more plausible by the apparent near coincidence of the outgassing and carbonate formation ages. The apparent presence of liquid water at later times, perhaps even up to the present day (Malin and Edgett, 2000), appears to be permissive of later carbonate formation also. Currently, however, there appears to be little rationale to consider alternate scenarios.

Although ALH84001 is a sample of the Noachian crust, it should not be considered a "typical" Martian crustal rock. It is an orthopyroxenite cumulate with much higher MgO and FeO, and lower Al_2O_3, SiO_2, and K_2O than typical Martian crustal rocks such as the Pathfinder "sulfur free rock" (Rieder *et al.*, 1997; Bell *et al.*, 2000). Although rocks at the Pathfinder site are thought to be derivative from the southern Martian highlands, the Al_2O_3 content of the "sulfur free rock" is much lower than that of lunar highland soils and even lower than in lunar mare soils. On average, the Martian crust appears to be rather Al_2O_3-poor. McSween and Keil (2000) conclude that if the global Martian dust is representative of the Martian upper crustal composition, the planet's surface geology is dominated by contributions from evaporitic salts and a basaltic protolith chemically similar to basaltic shergottites. Thus, fractionated MgO- and FeO-rich magmas may have been common within the crust, and orthopyroxenite ALH84001 may have formed as a mafic cumulate in a layered igneous province. The mode and timing of its formation may have been analogous to those of rocks of the lunar "Mg-suite", except that the composition of the surrounding crust was basaltic rather than anorthositic.

Finally, viewed from the context of the lunar samples, it seems fortuitous that even one out of 16 dated Martian rocks would have preserved such an old age. Only a few lunar crustal rocks have been reliably dated to have ages in the range of ~4.3–4.5 Gyr. Furthermore, the dated lunar ferroan anorthosites were small clasts extracted from lunar highland breccias, in which they often were surrounded by impact melt glass. The rarity of unadulterated "original" crustal material among the lunar highlands rocks poses some questions relative to ALH84001 and the Martian crust: Did Mars and Moon both experience the same heavy meteor bombardment early in their history? Did Mars experience a "terminal cataclysm" of bombardment? Was ALH84001 excavated from a considerable depth where it was shielded from bombardment? If 4.5 Gyr-old crustal rocks are fairly common on Mars, but not on the Moon, it may imply that the moon had a much heavier terminal bombardment that physically destroyed its crust. ALH84001 has given us a few clues, but answers to these questions await combination of spacecraft orbital imaging and absolute dating of samples returned from heavily cratered areas on Mars.

4.2. The Nakhlites and Chassigny: Amazonian or Hesperian Cumulates?

So far, we have no Martian meteorites that are clearly Hesperian in age; i.e., ~3.5–1.8 Gyr old, according to the Hartmann-Tanaka (HT) cratering model, or even older according to Hartmann and Neukum (2001). The ~1.3 Gyr radiometric ages of four cumulate rocks, three clinopyroxenite nakhlites (Nakhla, Lafayette, and Governador Valadares), and the dunite Chassigny (Figure 2) are Early Amazonian in the HT model, but close to the lower age limit of the Hesperian. However, they are squarely in the Middle Amazonian in the Neukum-Wise (NW) model. Close agreement of the ages of these four meteorites by four dating techniques, Rb-Sr, Sm-Nd, ^{39}Ar-^{40}Ar, and U-Pb, apparently unambiguously define their crystallization ages at ~1.3 Gyr (Figure 4). Type localities for the Early and Middle Amazonian are Amazonis Planitia (EA) and Acidalia Planitia (MA), respectively (Tanaka, 1986).

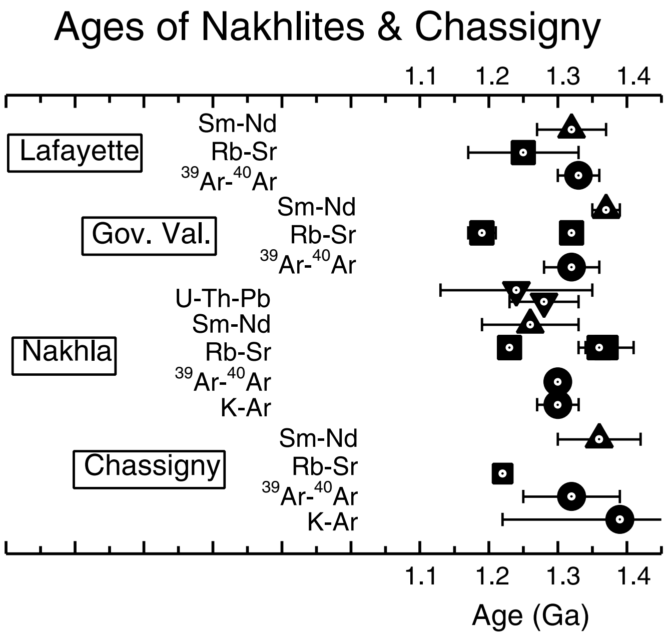

Figure 4. Radiometric ages of the Nakhlites and Chassigny. Nearly all the ages are compatible with an average age of ~1.3 Gyr.

Although the radiometric ages of the nakhlites are in general well-defined, they do show some isotopic disturbances, first evident in the Sr-isotopic heterogeneity noted by Papanastassiou and Wasserburg (1974). Some of the disturbance may be attributed to iddingsite, an apparent Martian weathering product formed during alteration by water. Some of the isotopic heterogeneity may perhaps be magmatic in origin. Attempts to date the formation time of iddingsite in the Lafayette nakhlite by the K-Ar and Rb-Sr techniques have yielded apparent ages of 600–700 Myr (Swindle *et al.*, 1999; Shih *et al.*, 1998), suggesting liquid water activity on Mars ~ 650 Myr ago. Malin and Edgett (2000) have cited a number of lines of evidence, such as the observations of gullies within the walls of a small number of impact craters, as indicating groundwater seepage and surface run-off on even younger Martian landforms. Examples are shown in this book (Hartmann, 2001).

4.3. THE SHERGOTTITES: LATE AMAZONIAN VOLCANISM

The preferred radiometric ages of basaltic and lherzolitic shergottites lie in the range ~165–475 Myr. Individual ages from the literature and unpublished data from the JSC lab are given in Table III and are shown with error limits in Figure 5. Confusion about the ages of the shergottites is slowly being dispelled. The initial apparent age discordance of the shergottites has been shown to arise from three sources: a) The presence of trapped ^{40}Ar from the Martian atmosphere and mantle, and possibly excess, inherited, radiogenic ^{40}Ar as well, in sufficient quan-

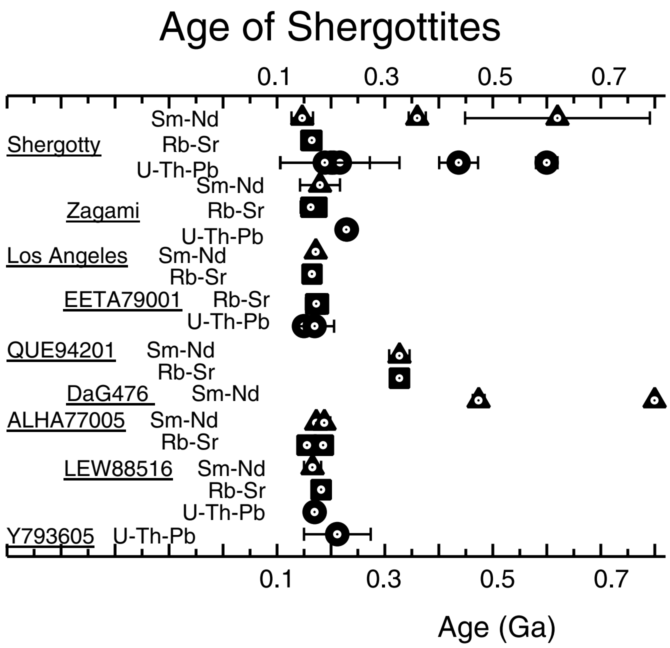

Figure 5. Radiometric ages of the shergottites. The Shergotty ages >0.3 Gyr seem to be in error due to unexplained analytical effects. Concordant Rb-Sr and Sm-Nd ages of ~327 Myr indicate that age is the true crystallization of QUE94201. The ubiquitous presence of terrestrial contamination has prevented determination of a Rb-Sr age for DaG476. The ubiquity of such contamination causes us to favor the 475 ± 11 Myr Sm-Nd age of Borg *et al.* (2000) to the older ~800 Myr age reported by Jagoutz *et al.* (1999) and Jagoutz and Jotter (1999) for this meteorite. See Table II for references.

tities to significantly affect measured ^{39}Ar-^{40}Ar ages. b) Analytical difficulties accompanying isotopic analyses of young samples with low abundances of the trace elements being analysed. c) The apparent presence of isotopic heterogeneity probably preserved in the basalts in the cores of pyroxene and olivine phenocrysts. These difficulties are being worked out, and preferred ages can be given with a degree of confidence (Table III).

It is also worth noting that a major contributor to initial confusion about the ages of the shergottites, and indeed all the SNC meteorites, was simply an early reluctance of meteoriticists to accept radiometric ages significantly less than ~4.5 Gyr as giving the time of igneous crystallization of any meteorite. The first radiometric age for a meteorite now considered to be of Martian origin was the K-Ar age of 580 ±50 Myr (recalculated with the decay parameters by Steiger and Jäger, 1977) determined for Shergotty by Geiss and Hess (1958) in an early study of the K-Ar ages of stony meteorites. Geiss and Hess (1958) considered their age to be "too young". They excluded the possibilities of K contamination or heterogeneity in the K content of the meteorite. Not knowing of Shergotty's Martian origin and, hence, assuming that Ar extracted from the meteorite could only consist of the radiogenic, spallogenic, and (terrestrial) atmospheric components, they considered Ar loss as being unlikely to explain its young age. A loss of ~95% of the Ar would be required, and they concluded that diffusive loss of that magnitude would be unlikely either via solar heating at the earth's orbit or beyond, or via frictional heating during passage through earth's atmosphere. We now know that correction for the "atmospheric" component using $^{40}Ar/^{36}Ar$ ~2000 for the Martian atmosphere, instead of 296 for the terrestrial atmosphere, gives an even younger age, more in agreement with currently accepted values for the Martian meteorites. Of course, the possible presence of a second atmospheric component, and the possibility that these young rocks might contain a mantle Ar component as well, introduces ambiguity in correcting for non-radiogenic ^{40} Ar. Later work showed that K-Ar ages <1 Gyr were relatively common among shocked chondritic meteorites and are produced by impact heating on meteorite parent bodies.

The question of the age of Shergotty was revisited by Bogard *et al.* (1979), who redetermined the K-Ar age using the ^{39}Ar-^{40}Ar technique, and also by Nyquist *et al.* (1979b), who determined the Rb-Sr age as well. The ^{39}Ar-^{40}Ar ages of stepped extractions of Ar from a whole rock were variable, but similar to the value of Geiss and Hess (1958). The ^{39}Ar-^{40}Ar ages of stepped extractions from a plagioclase separate were approximately constant and gave a good age plateau at 254 ± 10 Myr. The Rb-Sr age was younger still at 165 ± 11 Myr. Nyquist *et al.* (1979b) interpreted the ~165 Myr Rb-Sr age of Shergotty as due to metamorphic resetting because it was the lower of the Rb-Sr and ^{39}Ar-^{40}Ar ages. Subsequent work showed that not only was the Rb-Sr age younger than the ^{39}Ar-^{40}Ar age of 254 ± 10 Myr, but also younger than the still older Sm-Nd and U-Pb ages of Shih *et al.* (1982), Jagoutz and Wänke (1986), and Chen and Wasserburg (1986) (Table III; Figure 5). These latter measurements mostly have not been repeated, but it now seems likely that they were influenced by sample contamination, or other sources of analytical errors.

A recent *in situ* isotopic analysis by ion probe of U-Th-Pb in phosphates in Shergotty (Sano *et al.*, 2000) yields the time of closure of the U-Pb system in Shergotty phosphates as 204 ± 68 Myr ago. Whether this result excludes those Sm-Nd and U-Pb "ages" in excess of 300 Myr in Figure 5 as real crystallization ages depends on the actual mode of petrogenesis of these rocks. Jagoutz and Wänke (1986) preferred the older Sm-Nd age of 360 ± 16 Myr obtained from a pyroxene-leachate isochron to the younger Sm-Nd age of 147 ± 20 Myr they obtained from an isochron including the whole rock data. However, the lower value of ~147 Myr normally would be favored by the isotopic systematics, since data for the bulk ("whole rock") sample must lie on the isochron for closed-system evolution of the Sm-Nd system. Jagoutz and Wänke (1986) argued for an open Sm-Nd system, and that the phosphates crystallized from a metasomatic contaminant infiltrating a pyroxene cumulate. This possibility continues to be allowed by the results of Sano *et al.* (2000). Thus, the interpretation of Jagoutz and Wänke (1986) probably cannot be totally excluded, but it is nevertheless weakened by several observations. First, it requires Sr in plagioclase, as well as Nd in phosphates, to be derived from the metasomatic contaminant. Second, the Sm-Nd data for Shergotty also can be explained by recent terrestrial contamination, which may have been a greater problem than previously recognized. Third, Bogard and Garrison (1999) decomposed the Ar released from an irradiated sample of Shergotty into a radiogenic component that would be produced in 165 Myr of decay, and a trapped component with an $^{40}Ar/^{36}Ar$ ratio of 1780. This $^{40}Ar/^{36}Ar$ ratio is within the range of values believed representative of the Martian atmosphere, supporting the validity of this approach. Because most of the K and radiogenic $^{40}Ar^{*}$ is contained in plagioclase, this result would require that most of the K, as well as Nd and Sr, be attributed to the hypothesized metasomatic fluid. These more recent observations substantially support the earlier arguments of Jones (1986) against metamorphic resetting of the Rb-Sr ages of Shergotty and other shergottites by either thermal or hydrothermal events. Finally, a Sm-Nd age of 163 $\pm$ 7 Myr was found for Zagami, a "twin" of Shergotty (Nyquist *et al.*, 1995), concordant with three determinations of the Rb-Sr age averaging 177 $\pm$ 3 Myr. These arguments suggest that an age of crystallization as old as $\sim$360 Myr for Shergotty is unlikely. Our preferred age of 165 $\pm$ 4 Myr is the weighted average of the two Rb-Sr ages (Nyquist *et al.*, 1979b; Jagoutz and Wänke, 1986) converted to a ^{87}Sr decay constant $\lambda_{87} = 0.01402$ Gyr^{-1}, as used throughout this paper (Minster *et al.*, 1982).

That the young radiometric ages of the SNCs are indeed crystallization ages seems incontrovertible in light of the recent data summarized in Tables II and III. Three other basaltic shergottites, Zagami, EET79001, and Los Angeles have similar preferred ages as Shergotty; i.e., 177$\pm$3 Myr, 173$\pm$3 Myr, and 170$\pm$7 Myr, respectively. For Zagami, the value is based on three Rb-Sr ages and one Sm-Nd age (Table III). For EET79001, the preferred age is based on three Rb-Sr ages, one Sm-Nd age, and two U-Th-Pb ages. In this case the weighted mean is greatly influenced by a precise Rb-Sr age of 173 $\pm$ 3 Myr recently determined in the JSC lab (Nyquist *et al.*, 2001). For Los Angeles, the age is the weighted average of Rb-Sr and Sm-Nd ages. Although the preferred age for Shergotty appears to be slightly younger, its resolution from the other ages is problematic. The calculated uncertainty of $\pm$4 Myr for the Shergotty age may be unrealistically low in light of apparent cm-scale isotopic heterogeneity in Zagami (Nyquist *et al.*, 1995).

Two of the basaltic shergottites, QUE94201 and DaG 476, have older crystallization ages. Concordant Rb-Sr and Sm-Nd ages of 327 $\pm$ 12 and 327 $\pm$ 19 Myr were obtained for QUE94201 by Borg *et al.* (1997). QUE94201 contained easily-leachable components which may have formed as Martian weathering products. The presence of these phases and also of impact-produced glass veining throughout the rock led to significant complications of the isotopic systematics, but, nevertheless, the age appears to be robustly determined.

The age of DaG476, another basaltic shergottite with compositional similarities to QUE94201, is currently debated. Jagoutz *et al.* (1999), and Jagoutz and Jotter (2000) have presented Sm-Nd data leading to an apparent age of $\sim$800 Myr. Borg *et al.* (2000) found instead an age of 474 $\pm$ 11 Myr. This shergottite is heavily weathered, making the Rb-Sr data useless for age determination. Terrestrial contamination accompanies the weathering throughout the meteorite and is evident in elevated K and LREE abundances in some mineral phases (Crozaz and Wadhwa, 1999). Because the effect of terrestrial contamination would be to displace the apparent age to higher values, we favor the lower value of $\sim$474 Myr as most likely to be the true crystallization age.

Concordant Rb-Sr and Sm-Nd ages have been determined for two of the lherzolitic shergottites, ALH77005, and LEW88516. Our preferred ages for these two meteorites are 178 ± 6 Myr and 179 ± 6 Myr, respectively (Table III). An U-Pb age of $\sim$170 Myr for LEW88516 (Chen and Wasserburg, 1993) is also concordant with these values, as is an U-Pb age of 212 ± 62 Myr for the third lherzolitic shergottite, Yamato 793605 (Misawa *et al.*, 1997). These crystallization ages are the same, within uncertainties of a few percent, as the crystallization ages of several of the basaltic shergottites. Historically, these shergottites have been referred to as having crystallization ages of $\sim$180 Myr as suggested by Jones (1986), based on the 178 ± 3 Myr Rb-Sr age reported for Zagami by Shih *et al.* (1982).

The interpretation of the radiometric data for shergottites has been controversial in part because of the complexities that often exist both within and between radiometric systems. One of the most puzzling problems has been the observation that the apparent ^{39}Ar-^{40}Ar ages of the shergottites are systematically older than the Rb-Sr ages, as we have already mentioned for Shergotty. Although ^{39}Ar-^{40}Ar ages in terrestrial basalts can sometimes be "too old" because of inherited radiogenic ^{40}Ar, such situations are rare among meteorites. Bogard and Johnson (1983) first found that melt glass in the EET79001 shergottite contained trapped Martian atmospheric gases, including substantial amounts of ^{40}Ar, which had been shock-implanted by impacts on the Martian surface. Martian atmospheric gases have also been found in Zagami, ALH77005, and Y793605. In addition, an elementally fractionated component of the Martian atmosphere has been measured in some samples of the nakhlites and ALH84001 (see references in Bogard and Garrison, 1998). The ^{40}Ar/^{36}Ar ratio of the Martian atmosphere has a relatively high value of $\sim$1800 (Bogard and Garrison, 1999), which makes it difficult to correct for atmospheric ^{40}Ar using ^{36}Ar. Further, it is now recognized that some Martian meteorites contain a trapped volatile component from the Martian interior, which appreciably differs in elemental and isotopic composition from the atmospheric component (Ott, 1988; Bogard and Garrison, 1998; Marti and Matthew, 2000). The ^{40}Ar/^{36}Ar ratio of this interior component is not known and is probably variable. In many phases of Martian meteorites the Martian atmospheric and interior volatile components occur as mixtures in variable proportions.

Because of the presence of multiple Ar components, shergottites do not generally give reliable ^{39}Ar-^{40}Ar ages (Bogard and Garrison, 1999), as can be seen from the ^{39}Ar-^{40}Ar ages given for them in Table III. Most of the listed ^{39}Ar-^{40}Ar ages are from the compilation of Bogard and Garrison (1999), and are "total ^{40}Ar ages" for stepped ^{39}Ar-^{40}Ar analyses. Those ^{39}Ar-^{40}Ar ages that are closest to the preferred crystallization ages of the samples are for plagioclase separates of basaltic shergottites with relatively high modal abundance of plagioclase, and thus relatively high K-contents. Relatively good ^{39}Ar-^{40}Ar plateau ages of ∼254 Myr and ∼242 Myr were determined for Shergotty and Zagami feldspar. Also, as already mentioned, it was possible to decompose the ∼387 Myr total ^{40}Ar age for a bulk sample of Shergotty into trapped and radiogenic ^{40}Ar components. Data for Zagami feldspar are consistent with a similar decomposition of Ar components for a crystallization age of 180 Myr. However, Bogard and Garrison (1999) were unable to reliably determine the amount of radiogenic ^{40}Ar for other shergottite samples because of the multiplicity of Ar components that might have been present in the analyses. Terribilini *et al.* (1998) determined a conventional ^{40}K-^{40}Ar age of 196 ± 40 Myr from an isochron plot of ^{40}Ar/^{36}Ar versus K/^{36}Ar in a bulk sample and separated minerals of Shergotty. The corresponding ratio for trapped ^{40}Ar/^{36}Ar was found to be ∼1100, however, a value significantly lower than ^{40}Ar/^{36}Ar ∼1800 in the Martian atmosphere, showing that the shergottites contain both Martian atmospheric Ar and a mantle Ar component of significantly lower ^{40}Ar/^{36}Ar ratio. These two components mix in variable proportions in Martian meteorites, making it necessary to independently determine the ^{40}Ar/^{36}Ar ratio for each sample.

Subtle isotopic inconsistencies are present in the other isotopic systems as well. Some of these are manifest in the U-Pb ages summarized in Table III. Blichert-Toft *et al.* (1999) note also that Lu-Hf data for bulk shergottites do not show isochron relationships. Rb-Sr isochrons of different samples of the shergottites can give identical ages for different initial ^{87}Sr/^{86}Sr ratios (e.g., Nyquist *et al.*, 1995). Some isotopic inconsistencies for nakhlites probably are due to the presence of Martian weathering products. The differences in initial ^{87}Sr/^{86}Sr ratios among subsamples of shergottites and nakhlites require unusual petrogenetic processes.

As already mentioned, the Rb-Sr ages were originally interpreted as dating the time of shock metamorphism accompanying their ejection from Mars. An attraction of that explanation was that it accounted for the simultaneity of ages near 180 Myr. Nyquist *et al.* (1979b) argued that subsolidus *isotopic* equilibration might be achieved by heating at low temperatures (∼300 − 400 °C) for long times (∼10^4 yr) while allowing elemental zoning to be preserved. Jones (1986) criticized that interpretation on the grounds that preservation of elemental zoning patterns in major mineral phases of the shergottites, in spite of shock-induced transformation of plagioclase to maskelynite, precluded identification of the Rb-Sr ages with the time of shock metamorphism.

Because of the discordant ages obtained for Shergotty and other shergottites by the various methods, Shih *et al.* (1982) sought an approach that would "see through" secondary events, if such were the explanation of the young, ∼180 Myr ages. They noted that the whole rock Sm-Nd data for the basaltic shergottites, Shergotty and Zagami, combined with that of the lherzolitic shergottite, ALH77005, defined an apparent "isochron" of slope corresponding to an age of ∼1.34 Gyr, in remarkable agreement with the ages of the nakhlites. This coincidence was

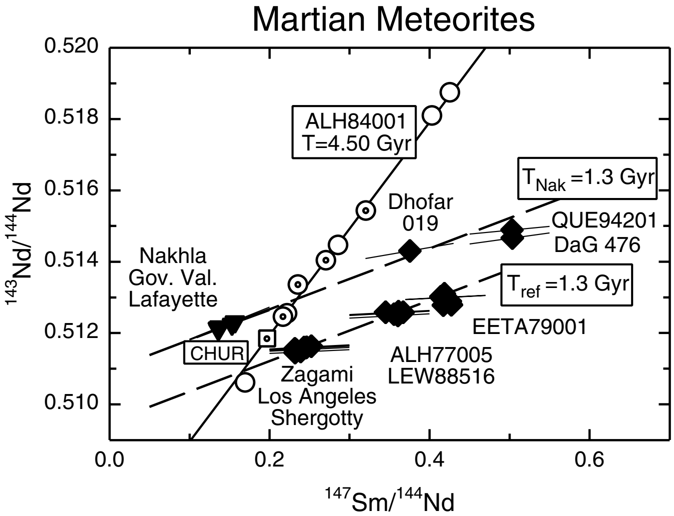

Figure 6. Whole rock Sm-Nd data for Martian meteorites. A 4.50 Gyr reference isochron is shown for bulk samples (*dotted circles*) and mineral separates (*open circles*) for ALH84001 orthopyroxenite. Dhofar019: preliminary data, Borg *et al.* (2001a). DaG476: constructed from the mineral isochron (Borg *et al.*, 2000) and the bulk ^{147}Sm/^{144}Nd ratio (Jagoutz *et al.*, 1999). The other data are from the literature. A reference 1.3 Gyr isochron (T_{ref}) has been drawn through the data for Shergotty, Zagami, and Los Angeles. For reasons that are unclear, the data of the shergottites Dhofar019, QUE94201, and DaG476 appear to lie along T_{nak}, a nakhlite isochron for the average age of the nakhlites Nakhla, Governador Valadares, and Lafayette. The linear alignment along T_{ref} from the traditional shergottites to EETA79001 has been interpreted as a mixing line between a mantle component to the right of EETA79001 and a "crustal" component to the left of the intersection of this line with the ~4.5 Gyr isochron. The isotopic data for a Chondritic Uniform Reservoir (CHUR) fall on this isochron (*squares*). Short lines through the individual data points show the slopes of mineral isochrons determined for these rocks. Rocks satisfying a simple two-stage isotopic evolution history would be derived from Martian mantle source regions having ^{147}Sm/^{144}Nd ratios determined by the intersection of the mineral isochrons with the primary ~4.5 Gyr mantle differentiation isochron. Such an intersection would occur at ^{147}Sm/^{144}Nd ~0.3 for QUE94201, DaG476, and Dhofar 019. QUE94201, DaG476, and the nakhlites also have measured excesses of ^{142}Nd from decay of 103 Myr ^{146}Sm (Harper *et al.*, 1995; Borg *et al.*, 1997; Jagoutz and Jotter, 2000), showing that Martian differentiation occurred very early. The likely addition of a crustal component to the parental magmas of the younger, ~175 Myr old shergottites would have displaced the apparent mantle ^{147}Sm/^{144}Nd ratios to the lower left along the primary mantle isochron.

reinforced by later data for basaltic shergottite EET79001. It seemed to suggest that the shergottites and the nakhlites might be related, if only indirectly, via mantle processes. It also suggested that the ~1.3 Gyr age might have significance for the shergottites. However, later interpretations have favored the view that this "isochron" is a mixing line, representing mixing among "crustal" and "mantle" end-members as suggested by Jones (1989) and Longhi (1991).

Figure 6 shows the currently available whole rock Sm-Nd data for Martian meteorites. Whole rock analyses and mineral separates for ALH84001 define an

$\sim$4.5 Gyr reference isochron, used here as a reference isochron giving the approximate age of the planet. The "traditional" basaltic shergottites (Shergotty, Zagami, and Los Angeles), plus the lherzolitic shergottites (ALH77005 and LEW88516) and basaltic shergottite EET79001 plot along an $\sim$1.3 Gyr reference isochron similar to the one originally defined for Shergotty, Zagami, and ALH77005 by Shih *et al.* (1982). The line in the figure is constrained to pass through the data for the traditional basaltic shergottites. According to the isotopic mixing models of Jones (1989) and Longhi (1991), the linear alignment of data is due to mixing of more radiogenic Nd (higher ^{147}Sm/^{144}Nd and ^{143}Nd/^{144}Nd ratios) from the Martian mantle with less radiogenic Nd (lower ^{147}Sm/^{144}Nd and ^{143}Nd/^{144}Nd ratios) from the Martian crust. End member compositions are to be found along the line to the right or left of the data array. This interpretation attributes no time significance to the linear data array, and suggests that the correspondence of the slope of the line to an apparent Sm-Nd age of $\sim$1.3 Gyr is a coincidence. This approach has much to recommend it. Norman (1999) found self-consistent results for the Nd isotopic compositions and REE abundances in Shergotty as a mixture between a LREE-depleted mantle-derived magma similar in composition to EET79001A and a LREE-enriched "crustal" component with $>$10 ppm Nd. The success of such models depends in large part on identification of at least one of the end member components. A number of meteorites have been suggested as representative of mantle-derived Martian magmas, including the nakhlites (Jones, 1989; Longhi, 1991), QUE94201 (Borg *et al.*, 1997), and EET79001 (Norman, 1999).

Recent data for newly found Martian meteorites have reopened the issue of whether Sm-Nd whole rock "isochrons" may have time significance, however. Jagoutz and Jotter (2000) note that leachates ($\sim$phosphates) from Nakhla, DaG476, and QUE94201 lie along a line of slope corresponding to an isochron age of 1.2 Gyr, and that these samples all have significant excess ^{142}Nd anomalies, showing that they all were derived from an early-formed mantle source with significant depletion in LREE relative to HREE (H for *H*eavy; *cf.* Harper *et al.*, 1995). Figure 6 shows that whole rock data for the nakhlites (Shih *et al.*, 1998, 1999), QUE94201 (Borg *et al.*, 1997), and newly found Dhofar019 (Borg *et al.*, 2001a) all plot close to the extension of the 1.3 Gyr nakhlite isochron. Because whole rock data for DaG476 are unreliable due to terrestrial contamination, we show a value calculated to lie on the mineral isochron of Borg *et al.* (2000) for the ^{147}Sm/^{144}Nd ratio measured by Jagoutz *et al.* (1999). If the $\sim$1.3 Gyr alignments have any time significance, it may be due to "mantle" end members generated by successive episodes of partial melting and magma extraction over an extended interval from before $\sim$1.3 Gyr ago until the crystallization ages of the basalts. Melt extraction on a massive scale may account for the large depletion of LREE abundances in QUE94201 and DaG476, and may have left surface evidence for large expanses of volcanic flows $\sim$1.3 Gyr ago. Models for the Nd and Sr isotopic evolution of QUE94201 (Borg *et al.*, 1997), however, require successive melt extractions from the source to be restricted to times near the basalt crystallization age.

In contrast to the Sm-Nd data, the whole rock Rb-Sr age for most Martian meteorites is ~4.5 Gyr (Shih *et al.*, 1982). Thus, the Rb-Sr data require planet-wide differentiation at ~4.5 Gyr, accompanied by establishment of mantle reservoirs that remained essentially closed systems to Rb/Sr fractionation thereafter. That major fractionation in Sm-Nd could occur $<\sim$1.3 Gyr ago without some additional fractionation of Rb/Sr is a strong constraint on isotopic models for crust/mantle differentiation. The models of Jones (1989) and Longhi (1991) for the petrogenesis of SNC meteorites invoked crustal assimilation following partial melting of the Martian mantle for nakhlite genesis at 1.3 Gyr ago and shergottite genesis at 180 Myr ago. Borg *et al.* (1997) presented a related multi-stage model of mantle melting and crustal assimilation for petrogenesis of QUE94201 at 330 Myr ago.

5. Ejection Ages and Events

Assuming a meteorite is ejected as a small object from the Martian surface and comes directly to Earth without secondary breakup in space, its "ejection age", i.e. the time since its ejection, equals the sum of its cosmic ray exposure (CRE) age and its terrestrial age (Eugster *et al.*, 1997b). Table IV gives all presently available CRE and terrestrial ages of Martian meteorites, and its notes explain how these ages were calculated. The CRE ages based on the stable noble gas isotopes ^{3}He, ^{21}Ne, and ^{38}Ar and appropriate production rates, T_s, may be subject to several sources of systematic bias, such as diffusive loss of spallogenic noble gases, which is possibly related to particular orbital parameters, errors in correction of production rates for variations in chemical composition, variations in shielding from cosmic ray particles, and contributions of solar particles to ^{21}Ne production. However, the ^{81}Kr-Kr ages, T_{81}, and, to a lesser extent, the ^{10}Be-^{21}Ne ages, T_{10}, are less subject to all these sources of bias (Marti, 1967; Eugster *et al.*, 1967). Terribilini *et al.* (2000) obtained T_{81} ages for 7 Martian meteorites. Their averages are 3% lower than the average T_s ages. The T_{10} ages, calculated for 5 Martian meteorites by several authors, are on an average also 3% lower than the T_s ages. It appears that these T_s ages have little or none of the biases mentioned above.

Figure 7a plots peak shock pressures from Table I versus ejection ages from Table IV. Both of these parameters are expected to be related to ejection of the meteorites from Mars. Most of the data, with exception of that for the nakhlites, plot within a range of shock pressures from ~30 GPa to ~80 GPa corresponding, respectively, to the pressures at which plagioclase is converted to maskelynite (P_{mask}), and at which shock melting occurs (P_{melt}). Melosh (1985) notes that in the spallation model for meteorite ejection, the ejection velocity is proportional to the pressure *gradient*, not the pressure. The implication of Figure 7 is that just prior to ejection, the shock pressure due to a compressive pulse at the shergottite location had reached nearly the maximum sustainable without shock melting. Furthermore, in the context of the Melosh model, the pressure gradient due to interfering com-

TABLE IV

Cosmic-ray exposure ages, terrestrial ages, and ejection ages of Martian meteorites in Myr.

	T_3	T_{21}	T_{38}	Meth.	$T_{s,av}$	T_{81}	T_{10}	T_{pref}	T_{terr}	T_{ej}
Shergottites (basalts)										
Dhofar019	(13.7[δ])	18.1[δ]	21.4[δ]	A	19.8 ±2.3			19.8 ±2.3		19.8 ±2.3
Los Angeles	(1.9[a])	3.0[a]	2.8[a]	A	3.02 ±0.30	3.10[b] ±0.70	3.10[β] ±0.20	3.10 ±0.20		3.10 ±0.20
	(1.3[γ])	3.2[γ]	3.1[γ]							
QUE94201	2.17[c]	3.12[c]	2.56[c]	A	2.50 ±0.25	2.10[b] ±0.25	2.6[o] ±0.5	2.42 ±0.20	0.29[o] ±0.05	2.71 ±0.20
	2.20[d]	2.60[d]	2.50[d]	A						
	1.91[e]	*	2.75[e]	A						
	2.0[f]	3.4[f]	2.3[f]	A						
Sayh al Uhaymir		1.5[ϵ]		A	1.5 ±0.3			1.5 ±0.3		1.5 ±0.3
Shergotty	2.56[g]	3.56[g]	2.54[g]	A	2.91 ±0.25	2.71[b] ±0.30	2.1[p] ±0.2	2.73 ±0.2		2.73 ±0.20
px	3.31[g]	3.17[g]	2.71[g]	A						
mask		3.9[g]	3.00[g]	A						
	2.4[h]	3.2[h]	2.3[h]	B		2.7[q] ±0.9				
	2.7[i]	3.9[i]	2.1[i]	B						
	2.5[f]	3.3[f]	2.2[f]	A						
	2.9[j]			B						
	3.2[k]	4.0[k]	2.7[k]	B						
	2.0[l]	3.4[l]	2.4[l]	B						
Zagami	2.84[c]	3.31[c]	2.56[c]	A	2.85 ±0.45	3.05[b] ±0.30		2.92 ±0.15		2.92 ±0.15
	3.2[k]	3.8[k]	2.6[k]	B						
	3.4[m]	4.3[m]	1.8[m]	B						
	2.1[n]	2.1[n]	2.2[n]	B						
Lherzolites										
ALHA 77005	3.64[t]	2.45[t]	2.50[t]	A	2.98 ±0.45		2.5[p] ±0.3	2.87 ±0.20	0.19[r] ±0.07	3.06 ±0.20
	3.8[m]	2.6[m]	2.9[m]	B			2.8[q] ±0.6		0.19[s] ±0.07 0.21[A] ±0.08	
LEW 88516	4.52[c]	4.90[c]	3.92[c]	A	4.01 ±0.40		3.0[w]	3.92 ±0.40	0.021[u] ±0.001	3.94 ±0.40
	4.42[x]	4.22[x]		B					0.021[v] ±0.001	
	3.96[z]	3.0[z]		B						
	4.4[w]	3.8[w]	3.0[w]	B						
Y793605	4.72[g]	3.98[g]	3.13[g]	A	4.67 ±0.50			4.67 ±0.50	0.035[A] ±0.035	4.70 ±0.50
	4.9[f]	5.2[f]	4.6[f]	A						
	5.36[B]	5.46[B]		A						

TABLE IV
(continued)

	T_3	T_{21}	T_{38}	M.	$T_{s,av}$	T_{81}	T_{10}	T_{pref}	T_{terr}	T_{ej}
Shergottite/Lherzolites										
DaG476	1.08[H]	1.26[H]	1.14[H]	A	1.16 ±0.11			1.16 ±0.11	0.085[α] ±0.050	1.24 ±0.12
DaG489	1.05[T]	1.36[T]	1.09[T]	A	1.17 ±0.19			1.17 ±0.19	0.085[α] ±0.050	1.25 ±0.20
EETA 79001 A	0.68[m]	0.52[m]	0.74[m]	B	0.54 ±0.09		0.78[L] ±0.14	0.60 ±0.09	0.012[K] ±0.002	0.73 ±0.15
	0.51[M]	0.43[M]	0.30[M]	B			0.5[P] ±0.1		0.320[L] ±0.170	
	0.61[L]	0.54[L]	0.45[L]	B			0.73[q] ±0.19		<0.06[P]	
		0.45[N]		B					(av.0.13) ±0.12	
EETA 79001 B	0.45[m]	0.69[m]	1.02[m]	B			0.90[q] ±0.17			
EETA 79001 C	0.46[M]	0.45[M]		B			$T_{10,av}$=0.73			
		0.42[N]		B						
Nakhlites										
Governador Valadares	12.2[P]	12.3[P]	6.7[P]	B	10.0 ±2.1			10.0 ±2.1		10.0 ±2.1
		9.5[Q]	9.2[Q]	B						
Lafayette	10.1[n]	12.4[n]	8.8[n]	B	11.9 ±2.2			11.9 ±2.2	0.0089[U] ±0.0013	11.9 ±2.2
	13.7[R]	16.0[R]	10.3[R]	B						
Nakhla	11.4[i]	12.4[i]	8.4[i]	B	12.2 ±1.5	10.75[b] ±0.40		10.75 ±0.40		10.75 ±0.40
	12.1[J]	12.0[J]	10.6[J]	B						
	14.8[R]	16.0[R]	12.2[n]	B						
Dunite										
Chassigny	13.3[g]	11.1[g]	10.5[g]	A	11.6 ±1.6	10.7[b] ±1.8		11.3 ±0.6		11.3 ±0.6
	14.7[F]	13.9[F]	7.1[F]	B						
	15.1[E]	13.7[E]	8.8[E]	B						
	12.3[i]	11.8[i]	7.1[i]	B						
Orthopyroxenite										
ALH 84001	15.4[ζ]	17.0[ζ]	12.3[ζ]	A	14.7 ±0.9	15.8[b] ±3.3		15.0 ±0.8	0.0065[v] ±0.0010	15.0 ±0.8
	15.6[c]	13.3[c]	12.2[c]	A						
opx	15.4[c]	15.4[c]	12.9[c]	A					~0.013[Y]	
	14.9[G]	17.7[G]	17.5[G]	A						
	15.3[t]	15.4[t]	12.8[t]	A						
	14.6[f]	14.1[f]	12.2[f]	A						

TABLE IV

(continued)

Notes

T_3, T_{21}, T_{38}: Cosmic-ray exposure ages based on ^{3}He, ^{21}Ne, ^{38}Ar and appropriate production rates.

Meth., M.: Method used for calculating T_3, T_{21}, T_{38}:

A – CRE ages as given by author(s). Applies to papers published after 1990.

B – CRE age calculated using production rates according to Eugster and Michel (1995). Applies to papers published before 1990 and to work where authors do not give CRE ages.

$T_{s,av}$: Mean value of all T_3, T_{21}, and T_{38} ages for a particular meteorite. Errors are $2\sigma_{mean}$.

T_{81}: ^{81}Kr-Kr CRE age (Terribilini *et al.*, 2000).

T_{10}: ^{10}Be-^{21}Ne CRE age as given by authors.

T_{pref}: Preferred CRE age, $T_{pref} = 0.5 \times [T_{s,av} + 0.5 \times (T_{81} + T_{10,av})]$.

Error of T_{pref}, $\Delta T_{pref} = 2\sqrt{\sum\left(2\Delta^2_{s,av} + \Delta^2_{81} + \Delta^2_{10,av}\right)/12}$.

T_{terr}: Terrestrial age as given by authors.

T_{ej}: Mars ejection age, $T_{ej} = T_{pref} + T_{terr}$. Error of T_{ej}, $\Delta T_{ej} = \sqrt{(\Delta T_{pref})^2 + (\Delta T_{ej})^2}$.

References

[a] Garrison and Bogard (2000); [b] Terribilini *et al.* (2000); [c] Eugster *et al.* (1997b); [d] Dreibus *et al.* (1996); [e] Swindle *et al.* (1996); [f] Garrison and Bogard (1998); [g] Terribilini *et al.* (1998); [h] Becker and Pepin (1986); [i] Ott (1988); [j] Eberhardt and Hess (1960); [k] Heymann *et al.* (1968); [l] Müller and Zähringer (1969); [m] Bogard *et al.* (1984); [n] Ott (1989), unpubl. data (see Schultz and Franke, 2000); [o] Nishiizumi and Caffee (1996); [p] Nishiizumi *et al.* (1986); [q] Pal *et al.* (1986); [r] Schultz and Freundel (1984); [s] Evans *et al.* (1992); [t] Miura *et al.* (1995); [u] Nishiizumi *et al.* (1992); [v] Jull *et al.* (1994); [w] Treiman *et al.* (1994); [x] Becker and Pepin (1993); [z] Ott and Löhr (1992); [A] Nishiizumi and Caffee (1997); [B] Nagao *et al.* (1997); [D] Bogard (1995); [E] Lancet and Lancet (1971); [F] Schultz and Signer (1973), unpublished; [G] Swindle *et al.* (1995); [H] Zipfel *et al.* (2000); [J] Stauffer (1962); [K] Jull and Donahue (1988); [L] Sarafin *et al.* (1985); [M] Becker and Pepin (1984), for R=35 to >1000 and d=6–80 g/cm^2; [N] Swindle *et al.* (1986); [P] Bogard and Husain (1977); [Q] Swindle *et al.* (1989); [R] Ganapathy and Anders (1969); [T] Folco *et al.* (1999), same production rates used as for DaG476; [U] Jull *et al.* (1997); [Y] Jull *et al.* (1995); [α] Nishiizumi *et al.* (1999; same T_{terr} for the paired meteorites DaG476/489); [β] ^{21}Ne and ^{10}Be data from Garrison and Bogard (2000) and Nishiizumi and Masarik (2000), respectively; [γ] Lorenzetti and Eugster (2001); [δ] Shukolyukov *et al.* (2000); [ϵ] Paetsch *et al.* (2000); [ζ] Eugster (1994).

pressive and tensile pulses, the latter reflected from the free surface of the planet, is sufficient to accelerate ejecta to escape velocity. These conditions define an annulus around the impact site from which material is ejected (*cf.* Warren, 1994).

From Figure 7a there appear to be 7 ejection events: at ~20, 15, 12, 4.5, 3, 1.3 and 0.7 Myr, respectively. Five of the total are for either basaltic or lherzolitic shergottites, suggesting that shergottites must be widespread on the Martian surface, or that the ejection mechanism preferentially selects basaltic compositions. Interestingly, both the oldest and the youngest events are for shergottites: Dhofar 019 at ~20 Myr ago, and EET79001 at ~0.7 Myr ago.

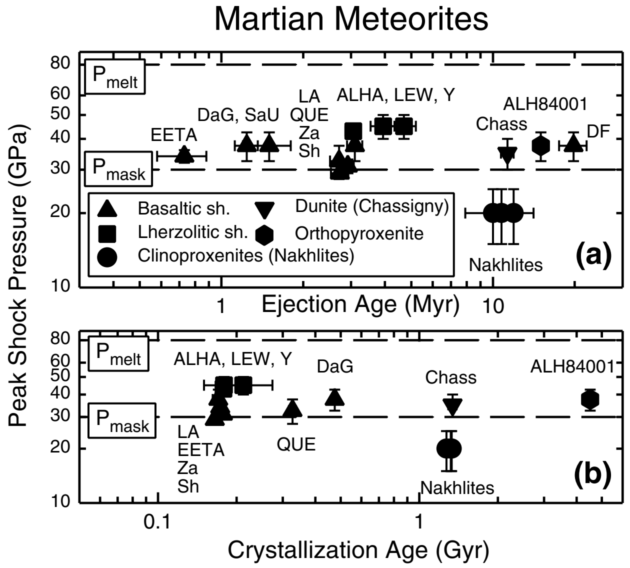

Figure 7. Peak shock pressures in Martian meteorites vs. their a) ejection ages and b) crystallization ages (EETA: EETA79001; DaG: Dar al Gani476; SaU: SaU005; LA: Los Angeles; QUE: QUE94201; Za: Zagami; Sh: Shergotty; ALHA: ALHA77005; LEW: LEW88516; Y: Y793605; Chass: Chassigny); DF: Dhofar019). Shock pressure data from Stöffler *et al.* (1986) and Stöffler (2000) except for SaU, LA, and DF (this paper; Table I); age data from Tables II, III, and IV (see references there). P_{max} and P_{melt} are the approximate peak shock pressures at which plagioclase is converted to maskelynite and basalts are shock-melted, respectively.

The picture changes somewhat when shock pressures are plotted against the crystallization ages of the meteorites (Figure 7b). To the extent that different Martian surface units are composed of rocks of distinct crystallization ages, both crystallization ages and ejection ages might be viewed as “event discriminators”. Nearly the entire Martian surface has been classified according to relative age as determined from the density of meteorite impact craters per unit area. These “crater retention ages” reflect ca. the upper 1 km of near-surface layers, so that lava flows of a variety of absolute ages may be present in a given area (Hartmann, 1999). Nevertheless, the crater retention ages provide a Martian context in which to view the potential number of ejection events. In this context, there appear to be only 4–5 events: one on old terrain, one on terrain of intermediate-to-young 1.3 Gyr age, 1 or 2 on young-to-intermediate, 0.3–0.5 Gyr terrain, and one on very young, ~0.18 Gyr terrain. The crystallization ages of Dhofar019 and SaU005 are not available yet, and might define additional events, possibly bringing the total to 6–7 separate events. More realistically, the total is apt to increase by no more than one, since SaU005 appears to be paired in ejection age with DaG476. Nevertheless, 7 meteorites, more than half of those for which crystallization ages have been determined, derive from ~0.18 Gyr terrain. This presents an apparent “paradox” of too many young meteorites from too little young Martian terrain (Nyquist *et al.*, 1998).

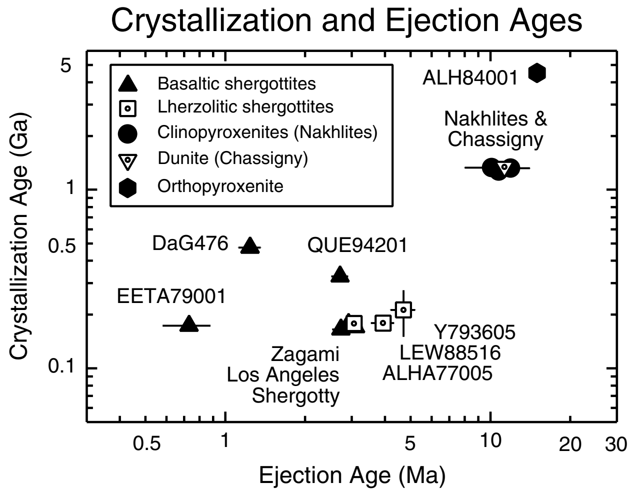

Figure 8. Preferred values of the crystallization ages of the Martian meteorites (Tables II and III) plotted vs. the ejection ages of the meteorites (Table IV).

Figure 7b seems to show that meteorites ejected in a given event may experience a variety of peak shock pressures, as shown for the "nakhlite–chassignite" and "young shergottite" events, respectively.

Figure 8 plots the crystallization ages of the Martian meteorites directly versus their ejection ages. SaU005 and Dhofar019 are two important omissions, particularly because the Dhofar019 event 19.8 ± 2.3 Myr ago on basaltic shergottite terrain is the oldest event of which we have record. The next event, which ejected ALH84001 from old (Noachian) terrain 15.0 ± 0.8 Myr ago, is clearly identified. ALH84001 apparently experienced a peak shock pressure of $\sim 35 - 40$ GPa on ejection, typical for Martian meteorites. This rock of initially deep-seated origin must have been previously moved to its place of ejection, probably a few meters to tens of meters beneath the Martian surface, assuming it came from just beneath the spall zone of an impactor ∼1–10 km in diameter (Melosh, 1984; 1985). This is consistent with its complicated shock history, consisting of two or more stages.

The next ejection event was the nakhlite event. The nakhlites, Governador Valadares, Lafayette, and Nakhla, were recovered at three widely separated localities: Egypt, Brazil, and Indiana. They are clinopyroxenites, have the same radiometric crystallization ages, and the same ejection age of ∼11.4 Myr. The dunite, Chassigny, is a fourth meteorite with a ∼1.3 Gyr crystallization age and an ejection age of 11.3 ± 0.6 Myr. Its radiogenic isotope composition and the abundances of several key incompatible trace elements suggest a close relationship to the nakhlites. Although Chassigny is mineralogically distinct from the nakhlites, it probably was ejected simultaneously with them. If so, it and the nakhlites probably were previously moved to their place of ejection. If, however, they crystallized near the center of a ∼100 m thick flow (Treiman, 1987), they may have been ejected directly by a large event. Alternatively, they may have experienced a two-stage shock history, with the earlier, lighter shock overprinted by the shock of the ejecting impact. In any case, early Amazonian Martian terrain, on the HT timescale, appears to have been sampled only once. We call this the NC event following Treiman (1995a).

The shergottites present the greastest puzzle. Taken at face value, the CRE ages indicate 4 ejection events within 5 Myr on terrain of three different, young, crystallization ages. Four basaltic shergottites, Shergotty, Zagami, Los Angeles, and QUE94201, have ejection ages in a narrow range of 2.7–3.1 Myr, and one lherzolitic shergottite, ALH77005, has a similar ejection age of 3.06 ± 0.20 Myr. The ejection ages of two other lherzolites (3.94 ± 0.40 Myr for LEW88516 and 4.70 ± 0.50 Myr for Y793605), however, do not overlap with those of the basaltic shergottites. The exposure ages of the three lherzolites are somewhat uncertain because of uncertainty in production rates, and because no ^{81}Kr ages yet exist. Further, ALH77005 and LEW88516 are similar in several properties (Treiman *et al.*, 1994) and might be expected to have the same ejection age. Thus, the differences in ejection ages among these three lherzolites are not easily interpreted.

Let us consider a scenario in which these 4 shergottites and the 3 lherzolites were ejected in a single event, and assume that the higher ages of LEW88516 and Y793605 are due to an exposure to cosmic rays prior to ejection from Mars (Nagao *et al.*, 1997). Nishiizumi and Caffee (1997) also concluded that for Y793605 it is not possible to completely eliminate, based on radionuclide activities, the possibility of a pre-irradiation on the surface of the parent body. On the other hand, Treiman *et al.* (1994) found from the radionuclide activities of LEW88516 that most of its cosmic-ray exposure probably occurred as an individual meteoroid. Furthermore, the cosmogenic ^{22}Ne/^{21}Ne ratios are 1.227 ± 0.035 for LEW88516 (Eugster *et al.*, 1997a) and 1.207 ± 0.020 for Y793605 (Terribilini *et al.*, 1998), indicating irradiation as small bodies for the total duration of exposure. A complicated exposure history beginning with excavation of LEW88516 and Y793605, followed by ejection within 1-2 Myr, or reburial of these two rocks between their initial excavation and ejection, seems improbable. Nevertheless, the shock features of Y793605 do suggest a two-phase shock history. These plutonic rocks probably crystallized at considerable depth, and were relocated to the ejection site. Peak shock pressures for all three are the highest among the Martian meteorites, suggesting that they were excavated from deeper depths than the other shergottites, and thus that they were more heavily shielded against cosmic-ray irradiation immediately prior to ejection. Nevertheless, the more brecciated texture of Y793605 suggests that it may have spent some time in a relatively shallow surface location. Thus, it may have acquired some cosmogenic noble gases between its first and second shock, the latter being coincident with its ejection from Mars. Nearly identical crystallization ages of 179 ± 5 Myr and 178 ± 8 Myr, respectively, for ALH77005 and LEW88516 (Table III), raise the possibility that all the lherzolites were ejected simultaneously with ALH77005, Shergotty, Zagami, and Los Angeles, even though the "ejection ages" of LEW88516 and Y793605 are analytically resolved from the others.

Additional events seem to be required for the shergottites/lherzolites DaG476 and 489, and EET79001. The ejection ages for EET79001, DaG476, and DaG489, are 0.73 ± 0.15 Myr, 1.24 ± 0.12 Myr, and 1.25 ± 0.20 Myr, respectively. DaG476 and 489 are likely to be part of a single fall, so the coincidence of their ejection ages is not surprising. The ages of DaG476/489 do not appear to overlap that of EET79001, and are clearly younger than those for the shergottites and the lherzolites. Zipfel *et al.* (2000) derive a slightly older ejection age of 1.35 ± 0.10 Myr for DaG476. In any event, its old crystallization age of $\sim$474 Myr indicates a separate ejection event for it. The ejection age of EET79001 has some ambiguity. Most CRE ages for EET79001 have been determined for lithology A, leading to a preferred value $T_{\text{pref}} = 0.60 \pm 0.09$ Myr (Table IV). This value, when combined with the averaged terrestrial age of 0.13 ± 0.12 Myr, yields $T_{\text{ej}} = 0.73 \pm 0.15$ Myr (Table IV). A significantly higher value of 1.22 ± 0.24 Myr would be obtained, however, by combining the ^{10}Be-^{21}Ne CRE age of 0.90 ± 0.17 Myr for EET79001 lithology B (Pal *et al.*, 1986) with the terrestrial age of 0.32 ± 0.17 Myr reported for EET79001 (Sarafin *et al.*, 1985). This is the highest value derivable from the EET79001 data and is in good agreement with the preferred ejection age of DaG476 (Table IV). Thus, the resolution of the EET79001 and DaG476 ages may not be completely established. The disparity in apparent ejection ages of lithologies A and B of EET79001 may be related to the chemical differences between them, which affect the spallation nuclide production rates used to calculate CRE ages.

6. Provenance of the Martian Meteorites

6.1. Alternative Ejection Scenarios

Inconsistent numbers of Martian ejection events are inferred from crystallization and CRE ages, respectively. One may argue, that either some of the CRE ages are due to secondary collisions after ejection from Mars, or that there are vast expanses on Mars that not only are "young", but are the same age, within current experimental uncertainty. In the following, we characterize these two approaches.

6.1.1. *Cosmic-Ray Exposure Events: Without Secondary Collisions*

This assumption takes the ejection ages at face value (Eugster *et al.*, 1997b). Figures 7a and 8 show the following events: S_{Dho} (Dhofar019 shergottite), O (orthopyroxenite), NC (nakhlites and Chassigny), S_L (lherzolitic shergottites, especially LEW88516 and Y793605), S (traditional basaltic shergottites), S_{DaG} (DaG476/489 and SaU005), and S_E (EET79001). These 7 events include 5 related to shergottites. Three of these are on terrain <500 Myr old. A modification of this scenario might allow an additional event, S_Q, for QUE94201 because of its unique crystallization age. Thus, in this scenario there are 7-8 ejection events, 5-6 of which are for shergottites $<\sim$500 Myr old, and three include shergottites $\sim$175 Myr old.

6.1.2. *Events Based On Crystallization Age: Cosmic-Ray Exposure With Secondary Collisions*

Nyquist *et al.* (1998) preferred the term "small body ages" to "ejection ages" to emphasize that the event directly dated was initiation of exposure of "small" bodies to cosmic radiation. "Small" was defined by Bogard *et al.* (1984) as $<\sim$6 m as a safe upper limit. A practical definition of "small" is that size below which a two-stage exposure becomes detectable in the total radionuclide and stable spallation nuclide data. This size is difficult to quantify, but is apt to be somewhat smaller, perhaps $\sim$3-4 m in diameter. In a "strong" version of this model, some CRE ages may be unrelated to the actual ejection events. In this case, the minimum number of events suggested by the crystallization age data are: O, NC, and S (Nyquist *et al.*, 1998). An additional two events S_Q, and S_{DaG}, might be allowed because the crystallization ages of QUE94201 and DaG476 differ from $\sim$175 Myr, and another two events may be permissible because precise crystallization ages of Dhofar019 and SaU005 are not yet available. The whole rock Sm-Nd data (Figure 6), and initial Sr- and Nd-isotopic data (not shown) suggest that DaG476, QUE94201, and Dhofar 019 constitute a suite of samples separate from the $\sim$175 Myr shergottites. Treiman (1995a) considered a larger array of petrologic, chemical, and chronological data, and suggested two distinct sites of origin for the then known SNC meteorites, which did not include QUE94201, DaG476/489, or Dhofar019. Thus, the probable number of events in this scenario is $\geq$ 4: O, NC, S_1, and S_2, where subscripts 1 and 2 separate the $\sim$175 Myr shergottites from $\sim$300 $-$ 500 Myr shergottites. Subgroup S_2 may be further sub-divided into as many as three. The $\sim$3 Myr exposure age for QUE94201 perhaps should exclude it from S_2. Also, SaU 005 is almost certainly paired with DaG476/489 on the basis of their CRE age. Note that for 4-6 ejection events the "secondary collision" hypothesis is required to hold only in a "weak" form for the young exposure age of EET79001. In that case, ejection events would have occurred at $\sim$20 Myr, $\sim$15 Myr, $\sim$12 Myr, $\sim$3 Myr, and $\sim$1.3 Myr, respectively, assuming that some of the lherzolitic shergottites had some exposure on the Martian surface.

That half or more of the total number of Martian meteorite ejection events yielded shergottites $<\sim$500 Myr old continues to be a paradox. According to new spacecraft observations, very late lava flows appear to cover an area of $\sim 10^7$ km^2 (Keszthelyi *et al.*, 2000), $\sim$7% of the total area of Mars, or $\sim$12% of the volcanic surface area (Tanaka *et al.*, 1992). The previously recognized Late Amazonian volcanic surface was only $\sim 3.3 \times 10^6$ km^2 (Tanaka *et al.*, 1992), so the apparent over-representation of shergottite-ejection events compared to the total number of events and to the number of orthopyroxenite and nakhlite events persists. This shergottite age paradox (Nyquist *et al.*, 1998) is most pronounced for scenarios involving no secondary collisions in meteorite transit from Mars. That is, the probability that of 3–4 random events, at $\sim$4.3(?), $\sim$3.0, $\sim$1.3, and $\sim$0.7 Myr ago, 2–3 ejected meteorites having indistinguishable crystallization ages within an analytical precision of a few percent, would seem to be very low. If these are truly separate, random, events, there must be great expanses of ∼175 Myr lavas on the Martian surface. Keszthelyi *et al.* (2000) noted that the $\sim 10^7$ km^2 of newly recognized flood basalts in Elysium and Amazonis Planitia is an area roughly the size of Canada. The age paradox appears to require an even larger area; i.e., ∼50% of the 84 × 10^6 km^2 total volcanic surface area of Mars (Tanaka *et al.*, 1992) appears to be required to be <∼500 Myr old, and ∼2/3–3/4 of that to be ∼175 Myr basalt if the events at ∼4.3, ∼3.0, and ∼0.7 are due to separate, random impacts on the Martian surface. The fewer the number of shergottite ejection events, the more easily reconcilable is their number with the Martian cratering record, and the number of nakhlite and orthopyroxenite ejection events (one each).

Finally, it should be noted that Martian surface ages are derived by scaling an assumed Martian cratering rate as a function of time to the lunar cratering rate as a function of the absolute ages of lunar surfaces as determined on returned lunar samples. Hartmann (1999) estimates the uncertainty in this process to be on the order of a factor of three. If the Martian cratering rate were higher than assumed in Tanaka *et al.* (1992), the apparent ages of Martian surface units would be shifted to lower values. Surface terrain of age ∼1.3 Gyr, assumed to be present in the Early Amazonian on the HT model, and representing ∼9% of the volcanic surface of Mars, would actually be found in the Late Hesperian, representing ∼14% of the volcanic surface. Surface terrain of ∼0.5 Gyr or less, found in the Middle and late Amazonian in the HT model, would be shifted to Early Amazonian, representing ∼15% of the volcanic surface. Together they would represent a sizeable fraction of the total surface (∼29%), and a significant probability that Martian meteorites would be ≤1.3 Gyr old. Even in this case, however, the relative frequency of shergottite events to nakhlite events would be expected to be only ∼1:1.

Mars may have been more active in recent times than previously thought, and late, thin lava flows may be relatively ubiquitous on the Martian surface, increasing the probability of young meteorites. Clarification of whether the ages of the meteorites are in proportion to the exposed area of young surfaces, however, will require a) continued and refined chronological studies of the Martian meteorites; b) absolute calibration of the Martian crater-frequency curve via returned Martian samples; c) more high resolution imagery of the Martian surface; and d) evaluation of the compositional biases, if any, that may exist in the yield of meteorites from surfaces of different compositional types.

6.2. Potential Source Terrains

Several authors have sought to identify potential Martian source terrains and candidate source craters for the Martian meteorites (Nyquist, 1983; Mouginis-Mark *et al.*, 1992; Treiman, 1995a; Barlow, 1997; Nyquist *et al.*, 1998). Such attempts have relied primarily on the crystallization ages of the meteorites and the perceived characteristics of the resulting impact craters. Mouginis-Mark *et al.* (1992) identified 9 candidate source craters for the SNC meteorites on the Tharsis plains of

Mars, the only area of the planet seen to have lava flows $\leq$ 1.3 Gyr old on the HT model using Viking imagery. The higher resolution imagery from the Mars Global Surveyor (MGS) has shown the presence of thin, nearly craterless, lava flows on areas that are more heavily cratered at the Viking resolution, however (Hartmann, 1999; Hartmann and Berman, 2000). These new observations significantly extend the Martian surface area known to be covered by young basalts, as well as lower some cratering ages into the range of the ages of SNCs by either model. Crater densities on these young lavas are 10^{-3} to 10^{-2} times the densities on lunar maria of ages $\sim$3.5 Gyr, so that some Martian basalts appear to be as young as $\sim$10 Myr or less (Hartmann and Berman, 2000).

6.3. Implications of Ejection Models

Three recent developments loosen the restrictions on possible source terrains for the shergottites: a) The MGS data show young, thin, lava flows covering older flows in some localities, particularly in Elysium and Amazonis Planitia (*cf.* Keszthelyi *et al.*, 2000). b) The older crystallization ages of $\sim$330 Myr, and $\sim$475 Myr found for QUE94201 and DaG476, respectively, further extend the potential source terrains. c) The lower theoretical limiting crater size for meteorite ejection of $\geq$3 km (Head and Melosh, 2000) increases the number of candidate source craters, as compared to those $\geq$10 km (Mouginis-Mark *et al.*, 1992).

The decrease in crater size required for meteorite ejection according to the Melosh (1984) spallation model greatly increases the number of fragments potentially ejected from Mars. According to Mileikowsky *et al.* (2000) the number of fragments ejected from craters $\sim$13 km diameter on Mars with shock pressures $<$1 GPa over 4 Gyr is 9.5×10^{10}. They estimate the number ejected at $<$1 GPa is about 2% of the total. For craters of this size, the mean ejecta fragment diameter is $\sim$0.3 m. These values may be compared to $\sim 9.4 \times 10^{9}$ fragments of mean diameter $\sim$0.9 m and shock pressure $<$1 GPa for a 30 km diameter crater, and $\sim 2.0 \times 10^{8}$ fragments of mean diameter $\sim$3 m and shock pressure $<$1 GPa for a $\sim$100 km diameter crater. The maximum fragment diameter is estimated to be about four times the mean diameter. The spallation model thus favors a high proportion of small meteorites, with the relative abundance falling off by more than 2 orders of magnitude for an order of magnitude increase in size. Although there is a high proportion of small meteorites among the lunar meteorites, small meteorites do not seem to be preferred among Martian meteorites (*cf.* Warren, 1994).

The largest Martian meteorite, Nakhla, has a recovered mass of 40 kg, corresponding to a spherical meteoroid $\sim$0.3 m in diameter. This, of course, is a lower limit on the pre-atmospheric size of Nakhla. Although 0.3 m is not large in an absolute sense, it is $\sim$60 times larger in mass, or $\sim$4 times larger in radius, than the largest lunar meteorite (Warren, 1994), implying proportionally larger impactors. For the spallation model, Mileikowsky *et al.* (2000) give fragment size $l = 3 \times 10^{-4}\ L$, where L is the impactor diameter. Since this scales as $V_{\mathrm{ej}}^{-2/3}$, where V_{ej} is the ejection velocity, we expect l_{Mars}/l_{Moon} ~1.6, implying moderately larger Martian launch craters than lunar launch craters. A much more significant difference would be implied if a "proto-EET79001" ~3 m in diameter is required for later involvement in a secondary collision. A fragment this large would be at the upper size limit for a 30 km crater. A larger crater, ~100 km in diameter, is inferred from the spallation model if the mean fragment size is to be ~3 m.

From Gladman (1997; Figure 11), we estimate that a Martian meteorite ejected 4 Myr ago would on average spend ~0.15 Myr in the main asteroid belt, where the collision rate is relatively high. Wetherill (1988) gave the half-life, τ (Myr), against collision in the main belt as $\tau \sim 1.2r^{1/2}$ for meteoroids of radius r (cm). Thus, for $r = 150$ cm, τ is 15 Myr, and on average ~1% of objects launched 4 Myr ago would undergo secondary collisions in the main belt. If a large number of fragments were launched simultaneously, as implied by recovery of 3–7 individuals from the shergottite (S) event, then the probability that one of those individuals would have undergone a secondary collision would be ~3–7%. Thus, the possibility of secondary collisions in space after ejection of meterorites from Mars cannot be discounted *a priori*. As noted earlier in the paper, crystalline rock fragments from the ~25 km diameter Ries crater shocked to the range of the Martian shergottites do not exceed a size of ~0.5 m. Fragment size should scale as the product of dynamic tensile strength and projectile diameter (Melosh, 1985). An increase in this product of a factor of ~6 or more for the Martian source crater of the shergottites compared to the Ries would appear to be required to yield proto-EET79001 fragments.

Mileikowsky *et al.* (2000) have estimated the minimum crater size theoretically consistent with proto-EET79001s of ~3 m at ~30 km. For ~0.3 m fragments, their estimate of an ~13 km final crater diameter is in close agreement with ~15 km estimated by Warren (1994) for Nakhla. This is ~15 times larger than the largest source craters Warren estimates for the lunar meteorites on the same basis, an unexpected difference considering $l_{Mars}/l_{Moon} \sim 1.6$. Warren (1994) also estimates that a crater ~11 km in diameter is required to launch the largest shergottite, Zagami. Thus, the relatively large size of the Martian meteorites may indicate derivation from relatively large craters, in spite of the lowering of the theoretical limit of the Melosh (1984, 1985) model for the smallest crater required for launch.

6.3.1. *Implications of the Apparent Shock Pressure Launch Window*

The observation of an apparent "launch window" of peak shock pressures for the Martian meteorites may have implications for their launch mechanism and provenance. Early attempts to account for the existence of Martian meteorites sought the explanation in unusual, relatively rare, events because conventional cratering theory did not account for them. It was argued, for example, that ejection velocities could not exceed twice the "particle velocity" produced in a compressive wave at the peak shock pressures the meteorites had experienced. This maximum velocity could be achieved only at a free surface. Stöffler *et al.* (1986) gave the particle velocities according to the Hugoniot equation of state for the then known shergottites as $\sim$1.5 $-$ 2.0 km/s for peak shock pressures of 29 $-$ 43 GPa, corresponding to ejection velocities of 3 $-$ 4 km/s. The additional meteorites included here widen the "launch window" to $\sim$20 $-$ 45 GPa, corresponding to particle velocities of 0.8 $-$ 2.2 km/s for the range of materials involved. Thus, their free surface launch velocities are $\sim$1.6 $-$ 4.4 km/s, and are below Martian escape velocities.

Nyquist (1983) suggested that vapor drag might provide the needed acceleration, especially in "richocheting", very oblique angle, impacts. Numerical simulations supported the idea that oblique impacts could launch meteorites from planets (O'Keefe and Ahrens, 1986). Vickery (1986) considered a range of possibilities involving vapor phase acceleration, and concluded that if the proto-SNCs had diameters of a few centimeters or less, gas acceleration from a 30-km-diameter crater would be consistent with their ejection from Mars. Interestingly, the equivalent spherical radii of Martian meteorites corresponding to their recovered masses are in the range $\sim$2-30 cm. Larger craters would give higher final velocities. For example, at a launch position, x, one crater radius, r_p, from the point of impact, 0.5 m rocks would be accelerated by factors of $\sim$1.7 for a 30 km diameter crater and of $\sim$2.9 for a 130-km crater. Vickery concluded that these factors, combined with the distribution of spall velocities at $x/r_\mathrm{p} = 1$ according to the Melosh (1984) model, would not result in velocities exceeding the Martian escape velocity. In fact, the effect of gas drag at $x/r_\mathrm{p} \leq 0.4$ was to decelerate "ejecta" from the spall model. Thus, vapor drag did not appear to augment the number of meteorites ejected compared to those ejected by spallation alone. Nevertheless, when applied to possible free surface launch velocities of $\sim$3-4 km/s for the shergottites, these factors result in velocities $\geq$5 km/s, as required for launch from Mars. Thus, vapor drag should not be ignored as a potential acceleration mechanism.

The oblique impact hypothesis has the attractive feature that craters produced by this process have a distinctive "butterfly" morphology that allows them to be identified. Nyquist (1983) and Nyquist *et al.* (1998) identified some candidate oblique impact source craters for the SNC meteorites, whereas Barlow (1997) identified a candidate oblique impact source crater for the orthopyroxenite ALH84001. Recently, attention has fallen on the Chicxulub "K-T impact" crater as of probable oblique impact origin. Schultz and D'Hondt (1996) show a time-lapse photographic sequence of the evolution of an impact-generated vapor cloud as well as summarizing the evidence for Chicxulub as an oblique impact. Hydrocode modeling of Chicxulub (Pierazzo and Melosh, 1999) and of oblique impacts more generally (Pierazzo and Melosh, 2000) describe the development of downstream melt and vapor plumes in such impacts. If surface fragments of Mars were launched into such oblique impact vapor plumes at velocities close to the escape velocity, perhaps some would acquire the "boost" needed for launch. Kadano and Fugiwara (1996) experimentally verified that for nylon projectiles impacting Cu targets, solid ejecta fragments are accelerated to velocities approaching that of the expanding vapor cloud. Continued studies should show whether oblique impacts could have a significant role in launching Martian meteorites.

However, the spallation model of Melosh (1984) is most widely considered. Warren (1994) applied it in detail to the launch of both lunar and Martian meteorites. In general the model worked well for lunar meteorites, with some modification to account for the high abundance of regolith breccias and the shallow launch depths of the lunar meteorites. There are no regolith breccias among the Martian meteorites, however. In other respects, also, including a preponderance of small meteorites relative to larger ones, and a majority of unshocked meteorites relative to moderately to highly shocked ones, the spallation model matches the lunar meteorites better than the Martian meteorites.

In the spallation model, interference of a reflected rarefaction wave from the free planetary surface with the direct compressive wave reduces shock pressures in a zone of interference, where near-surface material can be ejected at high velocity without experiencing high compressive shock pressures. Spalls form at the boundary of the interference zone with a lower "free-field" zone. In the "hydrodynamic ejection model" (Melosh, 1984, 1985, 1989), the ejection velocity is given as

$$V_{\mathrm{ej}} \approx 2V_{\mathrm{p}}(r)\left[1+(s/d)^2\right]^{-1/2} \tag{1}$$

which reproduces Equation 5.5.3 of Melosh (1989). Here, $V_{\mathrm{p}}(r)$ is the particle velocity in the compressive shock at a distance $r = (s^2 + d^2)^{1/2}$ from the "equivalent center" of the impact. The horizontal distance s is measured across the surface from the impact site to the ejection site. In analogy to nuclear explosions, d is the "equivalent depth of burst." The particle velocity varies with distance according to

$$V_{\mathrm{p}}(r) \approx C_v\,(U/2)\,(a/r)^{1.87} \tag{2}$$

Here, U is the velocity of an impactor of mean radius a, and $C_v \sim 1$ is a "coupling constant." The equivalent depth of burst is given approximately by

$$d \approx 2a\,(\rho_{\mathrm{P}}/\rho_{\mathrm{t}}) \tag{3}$$

Here, ρ_{P} and ρ_{t} are the projectile and target densities.

For $s = 0$, Equation (1) appears to give the "velocity doubling rule" for free surfaces, i.e., ejection velocity equal to twice the particle velocity. However, particle velocities corresponding to the observed peak shock pressures would lead to ejection velocities of only ~1.6–4.4 km/s for the Martian meteorites, as mentioned earlier. Equation (1) can only lead to ejection velocities in excess of 5.0 km/s if the theoretical particle velocity of Equation (2) is decoupled from the empirical particle velocity inferred from the peak shock pressures experienced by the meteorites. In this model, this "decoupling" of theoretical and empirical values of particle velocity is due to the interference of the compression and rarefaction waves.

Launch acceleration is proportional to the upward component of the pressure gradient acting on a volume of material (*cf.*, Melosh, 1984, 1989). Let the peak shock pressure in the compressive wave be P_{cw}, that in the rarefaction wave P_{rw}, and the resultant pressure in the zone of interference be P_{res}. Because the compressive wave arrives first at a given point in the target, P_{res} rises from zero as the compressive wave approaches, attains a maximum value, $P_{max} < P_{cw}$ as the tensile wave arrives, falls to a value P_{ten}, limited by the tensile strength of the material, as the maximum of the of the compressive wave passes, and finally returns to zero as the tensile wave passes (*cf.*, Melosh, 1985, Figure 3). The empirical particle velocities corresponding to peak shock pressures registered in the meteorites are those corresponding to P_{res}, whereas the particle velocity V_p in Equation (1) corresponds to P_{cw} of the compressive wave. Thus, in the interference zone very near the free surface, the velocity-doubling rule no longer applies, and further, as noted by Melosh (1985), the ejection velocity "is exactly twice the particle velocity in the compression wave only at the (unphysical) point on the surface lying directly above the equivalent center."

The complete decoupling of theoretical and empirical particle velocities appears to apply for lunar, but not for the Martian meteorites, in which the observed shock levels are at least partially coupled to the ejection phenomenon (Figure 7). Only a boost in acceleration above that recorded in the Martian meteorites is needed for their ejection, accomplishable as part of the spallation process, or by another ejection mechanism instead of, or in addition to, the spallation. Melosh (1989, Figure 5.8) presents contours of peak shock pressure and ejection velocity for the spallation model as a function of depth in units of the projectile diameter. Ejection velocity contours are given in units of impact velocity, U, and pressure contours in units of $\rho_t U^2$. For a typical impact velocity $U = 15$ km/s and target density $\rho_t = 3.3$ g/cm^3, peak shock pressures of 15 GPa $\leq P_{max} \leq$ 45 GPa for Martian meteorites give $\sim 0.2 \leq (P_{max}/\rho_t U^2) \leq 0.6$, corresponding to depth to diameter ratios, d/a, in the range $\sim 0.1 - 0.3$ very near the impact site and just below the near surface interference zone. Ejection velocities lie in the range $0.2 \leq (V_{ej}/U) \leq 0.5$, i.e., 3.5 km/s $\leq V_{ej} \leq$ 7.5 km/s for $U = 15$ km/s. Thus, for impactors $\sim$1 km in diameter, the depth of origin of the Martian meteorites is implied to be $\sim 100 - 300$ m. Such impactors would produce final craters $\sim$13 km in diameter (Mileikowsky *et al.*, 2000).

One apparent discrepancy between the predictions of this model and observations of the Martian meteorites concerns the apparent absence of lightly shocked Martian meteorites from our collections. According to the model (e.g., Melosh, 1989, Equation 6.4.3), the mass m_{ej} of material ejected at velocities greater than V_{ej} and shocked to pressures less than P_{max} is

$$m_{ej}/m = 1.2\,(P_{max}/\rho_t c_L U)\left[1 - \left(2V_{ej}/U\right)^{1/3}\right] \tag{4}$$

where m is the projectile mass, and c_L is the seismic wave velocity of the target. The other quantities are as previously defined. For $c_L = 6$ km/s, $P_{max} =$ 15 GPa, and other quantities as above, this equation predicts $m_{ej}/m = 0.08$. Likewise, $m_{ej}/m = 0.23$ is predicted for $P_{max} = 45$ GPa. Thus, we would expect that approximately 1/3 of the meteorites would have been shocked to pressures <15 GPa, whereas none are observed. The seriousness of this discrepancy is unclear because the number of recovered meteorites may be too few to give reliable statistics. Indeed, agreement between observation and theory is significantly improved if the lower value of P_{max} is shifted to $\sim$20 GPa, so that the nakhlites populate a low-shock bin. In this case, the observed ratio of "low-shock" to "high-shock" meteorites becomes 1: 6 in comparison to the predicted value of 1: 2.3. Nevertheless, the apparent lack of lightly shocked Martian meteorites is puzzling.

Because the functional relation between P_{cw} and particle velocity V_p varies with composition, there could be a compositional bias in the material ejected. Also, because both maximum pressure and particle velocity in the compressive wave decrease approximately as the inverse square of the distance from the impact (Equation 2; Melosh, 1985, Figure 1), and the area of an annulus of interference increases with the square of the distance, the majority of the material ejected will be near the largest value of r and the smallest value of V_p consistent with acceleration to escape velocity. This effect may lead to relatively long delays between arrival of the compression and rarefaction waves, and, thus, to relatively high values of P_{res} for the Martian meteorites.

Strong coupling of compressive shock into ejected Martian meteorites implies that they come from very near the surface contour of the spall zone. The most likely place of origin of the meteorites is from within the zone of "Grady-Kipp fragments" lying just below the spall zone illustrated in Figure 11 of Melosh (1984). The locus of the spall contour is given by z_p in Equation (3) of Melosh (1985). Warren (1994) estimates the ratio of z_p to projectile radius r_p as $z_p/r_p \sim 0.2 - 0.4$, and $z_p \sim 50 - 100$ m for a $\sim$10 km diameter crater produced by an impactor of radius $r_p \sim 250$ m. These depths are consistent with those estimated above by comparing meteorite shock pressures to the predicted shock contours of Melosh (1989), and allow excavation of the basaltic shergottites and possibly even the nakhlites from their places of igneous crystallization. The textures of the lherzolitic shergottites show that they crystallized in a large volume (Harvey *et al.*, 1993), and thus at greater depths, and suggests they were relocated to the place where they were ejected. Because coherent, homogeneous material is required for build-up of maximal pressure gradients, highly brecciated material from the Martian highlands probably is discriminated against in the launch process. Thus, ALH84001, although previously shocked, probably was part of a larger coherent block prior to its ejection from Mars, and not part of the Martian regolith.

7. Concluding Remarks

Further study of the shock metamorphic histories of the Martian meteorites, combined with an improved quantitative understanding of the ejection mechanism(s), can make an important contribution to determining launch conditions and prelaunch sample locations. As additional meteorites are discovered, it will be particularly important to establish whether the "launch window" of peak shock pressures is maintained. Head and Melosh (2000) conclude that most Martian meteorites come from relatively small events of crater diameter ∼3 km, just large enough to eject candidate Martian meteorite material. Nevertheless, several characteristics of the meteorites, including a relatively high percentage of "large" meteorites, strong launch pairing, sampling of a limited number of events, pre-launch shielding from cosmic rays, and, possibly, secondary break-ups in space, favor larger events. Warren (1994) noted that in the context of the spallation model, Nakhla and Zagami, representatives of two main ejection groups of Martian meterorites, probabably came from craters ∼15 km and ∼11 km in diameter, respectively. The launch mechanism for the Martian meteorites is sufficiently uncertain that a number of possible mechanisms should continue to be evaluated, however.

Because the places of origin of the Martian meteorites are unknown, use of their ages for calibrating the cratering rate is distinctly limited. Nevertheless, the observation of young igneous crystallization ages among the meteorites, down to ∼165 Myr, shows that Martian volcanism continues essentially until the present day. Moreover, the observation of a high proportion of young ages suggests that Mars has been volcanically relatively active at recent times. The grouping of Martian meteorite ages around certain preferred values emphasizes the need to correct for "launch-pairing" among them, in contrast to the lunar case, where most individual meteorites appear to represent individual ejection events (Warren, 1994). We note that efforts to determine reliable launch-pairing of the meteorites will enable better interpretation of a variety of mineralogical, geochemical, and isotope geochemical data obtained for these rocks, currently our only samples of Mars. Finally, it must be noted that the degree of reliability currently achieved for radiometric dating of the Martian samples results from two decades of experience and improvements in laboratory techniques. Not all of the problems encountered in dating these samples have been analytical. Martian rocks appear to bear the record of a complex series of igneous and secondary processes, and the return of Martian samples to terrestrial laboratories will be required to answer the many remaining questions of Martian chronology. Those samples, too, will doubtless hold surprises for unwary analysts, but, if adequate samples are returned, dating them absolutely should be possible.

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Address for correspondence: SN/Planetary Sciences, NASA Johnson Space Center, 2101 NASA Road 1, Houston, TX 77058-3696, USA;
(laurence.e.nyquist1@jsc.nasa.gov)