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The rocks of Mars, from far and near

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Abstract—The age, structure, composition, and petrogenesis of the martian lithosphere have been constrained by spacecraft imagery and remote sensing. How well do martian meteorites conform to expectations derived from this geologic context? Both data sets indicate a thick, extensive igneous crust formed very early in the planet's history. The composition of the ancient crust is predominantly basaltic, possibly anesitic in part, with sediments derived from volcanic rocks. Later plume eruptions produced igneous centers like Tharsis, the composition of which cannot be determined because of spectral obscuration by dust. Martian meteorites (except Allan Hills 84001) are inferred to have come from volcanic flows in Tharsis or Elysium, and thus are not petrologically representative of most of the martian surface. Remote-sensing measurements cannot verify the fractional crystallization and assimilation that have been documented in meteorites, but subsurface magmatic processes are consistent with orbital imagery indicating thick crust and large, complex magma chambers beneath Tharsis volcanoes. Meteorite ejection ages are difficult to reconcile with plausible impact histories for Mars, and oversampling of young terrains suggests either that only coherent igneous rocks can survive the ejection process or that older surfaces cannot transmit the required shock waves. The mean density and moment of inertia calculated from spacecraft data are roughly consistent with the proportions and compositions of mantle and core estimated from martian meteorites. Thermal models predicting the absence of crustal recycling, and the chronology of the planetary magnetic field agree with conclusions from radiogenic isotopes and paleomagnetism in martian meteorites. However, lack of vigorous mantle convection, as inferred from meteorite geochemistry, seems inconsistent with their derivation from the Tharsis or Elysium plumes. Geological and meteoritic data provide conflicting information on the planet's volatile inventory and degassing history, but are apparently being reconciled in favor of a periodically wet Mars. Spacecraft measurements suggesting that rocks have been chemically weathered and have interacted with recycled saline groundwater are confirmed by weathering products and stable isotope fractionations in martian meteorites.

INTRODUCTION

Rocks are arguably the most informative kind of planetary samples. Unlike soils, fluids, and gases whose compositions reflect their time-integrated histories, rocks record discrete geologic processes and events that can usually be characterized from their mineralogy, geochemistry, and textures. Yet, more than two decades after the shergottite–nakhlite–Chassigny (SNC) meteorites were proposed to be martian rocks (McSween et al., 1979; McSween and Stolper, 1979; Walker et al., 1979; Wasson and Wetherill, 1979), they remain puzzling in many ways. This should not really be a surprise, given the geologic complexity of Mars revealed by spacecraft missions. Mars is unlikely to give up its secrets easily, and the emerging puzzle must be scrutinized repeatedly from different perspectives, as new pieces become available.

The SNC meteorites include volcanic or subvolcanic (basaltic shergottites and nakhlites) and plutonic (lherzolitic shergottites, Chassigny, and Allan Hills (ALH) 84001) rocks. Thorough petrologic, geochemical, and chronologic summaries of martian meteorites have been published elsewhere, for example, McSween and Treiman (1998); Meyer (1998), Nyquist et al. (2001), and references therein; augmented by descriptions of the newly recovered shergottites Los Angeles (Rubin et al., 2000; Mikouchi, 2001), Dar al Gani (DaG) 476/489/670/735 (Folco et al., 2000; Zipfel et al., 2000; Mikouchi et al., 2001), Sayh al Uhaymir (SaU) 005/008/051/094 (Zipfel, 2000; Hofmann et al., 2001), Dhofar 019 (Taylor et
Evidence that SNC meteorites are of martian origin has recently been reviewed by Treiman et al. (2000). The most telling evidence is the presence of trapped gases with elemental and isotopic compositions that match those of the apparently unique martian atmosphere. These gases were implanted during shock, probably by the impact events that ejected them from the planet’s surface. Other arguments are based on consistency between the chemical compositions and fractionations, oxidation states, and chronology of SNC meteorites and Mars. In some cases, these arguments point to formation on a large, geologically complex planet, but not specifically Mars.

A major challenge in using SNC meteorites as exploration tools is that we cannot yet fix their locations on the planet. However, remote-sensing data from Mars orbiters and landers can provide some geologic context. Several times previously (McSween, 1985, 1994) I have reviewed what we know, or think we know, about the geology of Mars based on SNC meteorites. Here we will invert that logic, by considering what can be construed about the rocks of Mars from the planet’s geology, as revealed by spacecraft, and then by examining how well martian meteorites conform to these expectations. The "far and near" of the paper’s title thus refer to vantage points afforded by spacecraft and by laboratory investigations.

AGE AND STRUCTURE OF THE MARTIAN CRUST

Geologic Context

The crust in the southern hemisphere of Mars is heavily cratered and thus ancient (assigned stratigraphically to the Noachian period, >3.5 Ga; Hartmann and Neukum, 2001). The northern hemisphere has been resurfaced, mostly during the subsequent Hesperian period. This crustal dichotomy is expressed topographically, as the elevation of the southern highlands is ~5 km higher than the northern lowlands (Smith et al., 1999). An inventory of depressions ("ghost craters") in the northern lowlands obtained from Mars Global Surveyor (MGS) laser altimetric data (Frey et al., 2001) reveals that the density of large craters beneath the northern plains is comparable to that of the southern highlands, suggesting that basement in the north is also Noachian in age. Giant impact basins, thought to have been formed during a period of late bombardment at ~4 Ga, occur in both the southern and northern hemispheres, for example, Hellas and Utopia, respectively. Although Hellas and Utopia have nearly the same diameter, Utopia is shallower by 8 km, which constrains the thickness of the northern lowlands fill in this area (Smith et al., 1999).

The Tharsis rise, which straddles the dichotomy boundary, contains huge shield volcanoes. MGS studies indicate that loading by Tharsis produced an encircling topographic trough and a ring of negative gravity anomalies (Phillips et al., 2001). Formation of the Tharsis bulge resulted in significant fracturing and faulting, expressed most prominently in the Valles Marineris canyon system. That Tharsis is a massive (3 × 10^8 km^3) accumulation of magmatic rocks is bolstered by MGS imagery which suggests that the full extent of the Valles Marineris canyon walls is a volcanic sequence (McEwen et al., 1999). Although Tharsis volcanoes and volcanic plains are not heavily cratered, indicating relatively young (Amazonian age, <2.9–3.3 Ga; Hartmann and Neukum, 2001) eruptions, this locus of volcanism has persisted for billions of years (Phillips et al., 2001; Wilson et al., 2001). Elysium in the northern lowlands is another volcanic rise of smaller scale.

Because no seismic measurements are available, the thickness of the martian crust is only constrained by MGS topography and gravity. Simple models, with uniform densities for crust and mantle but different assumptions about elasticity, give a lower limit of 40–50 km for the global crustal thickness (Nimmo and Stevenson, 2001; Zuber, 2001). Even this lower limit corresponds to >4% of the planetary volume, significantly greater than the Earth’s proportion of crust. The average crust is unlikely to be >100 km in thickness, but the crust underlying Tharsis must be this thick to provide isostatic support for its high-standing topography (Solomon and Head, 1982). Moreover, if bouyancy is the sole driving force for ascending magmas, the summits of large Tharsis volcanoes indicate sources at depths >120 km.

Age of the Crust from Martian Meteorites

The crystallization age of ALH 84001 at ~4.5 Ga ago (Nyquist et al., 2001) directly dates crust formation, and its 40Ar/39Ar age of ~4.0 Ga has been interpreted to reflect a late bombardment (Ash et al., 1996). The igneous and impact chronologies of this meteorite are thus consistent with its derivation from the heavily cratered southern highlands, although some Noachian basement may also be exposed in the northern hemisphere (e.g., Arabia Terra). The extraction of incompatible elements from the mantle during crust formation also affected several radiogenic isotope systems in other SNC meteorites. Early fractionation at ~4.5 Ga is clearly demonstrated by whole-rock 87Rb/87Sr isochrons (Borg et al., 1997) and the close fit of whole-rock lead isotopic data to the Pb-Pb geochron (Chen and Wasserburg, 1986). In addition, excess 142Nd, formed by rapid decay of now-extinct 146Sm,
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has been documented in some SNC meteorites (Harper et al., 1995; Borg et al., 1997). The existence of this isotopic anomaly requires rapid differentiation of mantle and crust prior to decay of the parent isotope. Isotopic evidence of early crust formation is consistent with spacecraft observations of pervasive outcrops and covered basement of Noachian age.

All SNC meteorites except ALH 84001 have relatively young crystallization ages (Fig. 1), which date igneous additions to the crust at specific sites. The nakhlites and Chassigny crystallized ~1.3 Ga ago, and the shergottites span an interval from 165 to 875 Ma with most samples clustering near the recent end of this range. The young ages of shergottites remain controversial. Nyquist et al. (2001), following Jones (1986), argued that the ~175 Ma ages represent times of crystallization, whereas Blichert-Toft et al. (1999) favored the interpretation of Jagoutz and Wänke (1986) that crystallization occurred earlier and the young ages result from fluid metasomatism. Inconsistencies between different isotopic systems of these meteorites relate to magmatic assimilation of isotopically distinct crust, partial resetting during shock, and other factors. I find the arguments for young ages reflecting crystallization compelling and, for the purposes of this paper, will accept these ages at face value. The crystallization ages of most basaltic and lherzolitic shergottites are identical, suggesting they formed in a common igneous event. Mittlefehldt et al. (1999) proposed that the basaltic shergottite Elephant Moraine (EET) A79001 formed as an impact melt, and other shergottites with similar petrography (the Dar al Gani, Sayh al Uhaymir, and possibly Dhofar meteorites) might conceivably share this origin. If true, their crystallization ages (~175, 475, and 575 Ma) define impact events and thus provide only minimum ages for parts of the crust. Based on crater-counting chronology, most of the crust having ages ≤1.3 Ga lies in Tharsis or Elysium (Mouginis-Mark et al., 1992a; Hartmann and Neukum, 2001). Even if we ignore the shergottites that might be impact melts, the range of crystallization ages for nakhlites and other shergottites still implies an extended period (>1 Ga) of magmatic activity, in conformity with estimates of the episodic lifetime for construction of large Tharsis volcanoes (Wilson et al., 2001).

![Diagram](image)

**Fig. 1.** Comparison of crystallization and ejection ages for martian meteorites, as summarized by Nyquist et al. (2001) with additional data for Dho019 (Borg et al., 2001; Park et al., 2001). Other meteorites with measured ejection ages but unknown crystallization ages are shergottites DaG 489 = 1.1 Ma (Park et al., 2001) and SaU 005 = 1.0 to 1.5 Ma (Patsch et al., 2000; Park et al., 2001), similar to DaG 476; shergottite NWA 480 = 2.4 Ma (Marty et al., 2001), similar to EETA79001; and nakhlite NWA 817 = 9.7 Ma (Marty et al., 2001), similar to other nakhlites.
Is SNC Sampling Chronologically Representative?

It is perplexing that so many martian meteorites are derived from young volcanic terrains that comprise such a small part of the planet's surface (Nyquist et al., 1998). Times of meteorite ejection can provide insight into the number of sites that have been sampled. Because the estimated orbital travel time from Mars to Earth is short (Gladman, 1997), the ejection time is taken as the sum of a meteorite's cosmic-ray exposure age and terrestrial residence age (Eugster et al., 1997). SNC ejection ages are compared with crystallization ages in Fig. 1. Clusters of ejection ages almost certainly represent discrete impact events at ~3 and ~11 Ma. Interpreting the ejection ages of other shergottites is problematic. Lherzolitic shergottite ALHA77005 was ejected at ~3 Ma, but two other petrologically indistinguishable lherzolites have ejection ages of ~4 Ma. Although this terrain might have been sampled twice by impacts a million years apart, that seems unlikely. The modest difference in ejection ages may reflect uncertainties in cosmogenic nuclide production rates due to differences in bulk composition (Garrison and Bogard, 1993), or pre-irradiation of some lherzolitic shergottites on the martian surface (Nyquist et al., 2001). EETA79001 and its Dar al Gani and Sayh al Uhaymir relatives have younger ejection ages of ~1 Ma. The ~1 Ma ejection cluster may represent a discrete impact event or break-up of a larger meteoroid in space. ALH 84001 (~15 Ma) and Dho far 019 (~15 Ma) apparently represent separate ejection events. When these meteorites are included, the SNC ejection ages represent as few as four or as many as six events spanning 14 Ma.

Several conclusions can be drawn from a comparison of crystallization ages and ejection ages (Fig. 1). The nakhlites and Chassigny that comprise the ~11 Ma cluster of ejection ages crystallized at the same time. In contrast, the ~3 Ma cluster contains volcanic and pluto nic shergottites that apparently crystallized over an interval of 150 Ma, and the ~1 Ma cluster contains meteorites with crystallization (or shock) ages that span 300 Ma. Because the ejection mechanism is understood to launch only spalls of material on or very close to the surface (Melosh, 1984), craters large enough to sample several volcanic units having different ages are required. The requirement for large impacts might be relaxed if the target surface were heterogeneous because of impact excavation and mixing. However, all martian meteorites have suffered shock at peak pressures of 15–45 GPa (Nyquist et al., 2001), probably reflecting a launch window created by fairly large impact events. If each ejection age or cluster is a discrete impact event (Eugster et al., 1997), the Amazonian surface (Tharsis or Elysium?) has been abundantly sampled. I favor fewer ejection events, but oversampling of young crust still persists. It appears that young, coherent igneous rocks are the only samples that routinely survive the ejection process (Warren, 2001) or that impact disaggregation of older terrains may hamper transmission of the intense shock waves required to eject near-surface spalls.

Volume of the Crust from Martian Meteorites

Martian meteorites do not contain primary high-pressure minerals that directly constrain crustal thickness. Sohn and Spohn (1997) derived a geophysical model with a considerably thicker (100–250 km) basaltic crust than estimates based on spacecraft data. The thick crust was inferred partly on an assumption that martian meteorites were products of extensive partial melting, but SNC compositions are better understood as reflecting extensive fractionation. Norman (1999) employed a mass-balance model for rare earth element (REE) abundances and Nd isotopic compositions of shergottites to estimate the volume of the crust. A globally averaged crustal thickness of ~45 km is indicated by the inferred trace element composition of crust that was assimilated by shergottite magmas. Norman's preferred thickness (20–30 km) is less than that derived from MGS topography and gravity, but his upper limit agrees with the lower limitation on thickness from these models. This estimate used EETA79001 to specify the mantle Nd abundance and isotopic composition; Queen Alexandra Range (QUE) 94201, probably a more appropriate choice for a mantle-derived magma, would yield a thicker crust of ~75 km. These estimates bracket the thickness estimate from spacecraft data.

COMPOSITION OF THE MARTIAN CRUST

Orbital Spectroscopy
Igneous Processes

Geologic Context

The martian surface consists of dark regions with low thermal inertia, interpreted as mixtures of rocks and sand-size particles, and bright red regions with high thermal inertia, which are covered by deposits of fine-grained dust. Mineralogically diagnostic visible and near-infrared (VISIR) spectral features in dusty regions are obscured by nanophase iron oxides, and the micron-size dust mantles produce black-body emittance in thermal infrared spectra. Consequently, spectral measurements from orbiting spacecraft reveal nothing about rock compositions in dusty regions, which unfortunately constitute about half the planet. Spectral measurements do provide information about the mineralogy of the dark regions, but it is worth noting that both VISIR and thermal emission spectra of low-albedo areas are dominated by sand-size particles rather than rocks.

Orbital mapping of dark regions of the martian surface using the MGS thermal emission spectrometer (TES) suggests that
basalt dominates the southern highlands and andesite is largely localized in the northern lowlands (Bandfield et al., 2000). Rock assignments are based on similarities between martian and laboratory spectra and on deconvolved mineralogy. More extensive calibration using a variety of volcanic rock types (Wyatt et al., 2001) confirms the spectral distinction between basalt and andesite and further shows that reasonably accurate bulk-rock chemistry can be calculated from the deconvolved TES mineralogy. Silica and alkali abundances, estimated from the mineralogy of the average basalt (surface 1) and andesite (surface 2) of Bandfield et al. (2000) are shown in Fig. 2 (Hamilton et al., 2001a). Based on this commonly used classification scheme, surface 1 actually corresponds to basaltic andesite, although this might reflect admixture of other materials with basalt.

The northern plains, which are the locus of materials interpreted to have andesitic compositions (Fig. 3), have also been suggested as the site of an ancient ocean basin (Head et al., 1999). This smooth, resurfaced area may be covered by sediments, as inferred from MGS gravity and topography (Zuber et al., 2000). Wyatt and McSween (2001) showed that TES spectra of Mars andesite and weathered basalt were virtually identical, and demonstrated that the andesite spectrum could be equally well deconvolved as a mixture of relic igneous minerals (plagioclase and pyroxene) and alteration minerals (mostly clays). They hypothesized that material identified as andesite may actually be basalt that experienced submarine weathering on a seafloor, or sediments derived from weathered basalts and deposited within this basin. The hypothesis that Mars is a basalt-covered world is bolstered by the basalt-like chemistry of martian soils (Clark et al., 1982; Rieder et al., 1997; McSween and Keil, 2000), which were presumably derived by weathering or comminution of common surface rocks.

Similarities in the Phobos–2 VSNIR spectra of Syrtis Major and that of basaltic shergottites previously led to suggestions that shergottite-like basalts may be common volcanic rocks on Mars (Mustard and Sunshine, 1995; Mustard et al., 1997). The spectra are dominated by pyroxene absorption bands, and spectral deconvolution indicates coexisting pigeonite and augite having compositions like the pyroxenes in shergottites. The deconvolved TES mineralogy shows

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**Fig. 2.** Chemical classification of martian volcanic rocks. The composition of basaltic materials in the southern highlands (surface 1) derived from deconvolved TES spectra (Hamilton et al., 2001a) is distinct from basaltic shergottites. However, the Mars Pathfinder dust-free rock composition (Wänke et al., 2001) is similar to the andesite (surface 2) composition derived from TES spectra (Hamilton et al., 2001a).
basaltic materials dominated by plagioclase and augite, but apparently without significant pigeonite (Christensen et al., 2000a; Hamilton et al., 2001a). Thus there is an apparent discrepancy in the mineralogy of basalts inferred from VIS-NIR and thermal emission spectra. Although olivine has been identified in the TES spectra of a few basalt localities on Mars (Hamilton et al., 2001b), it is relatively uncommon, at least above the detection level (~2.5%). Deconvolved modes from TES spectra of martian basalt and andesite have plagioclase > pyroxene (Hamilton et al., 2001a). Plagioclase compositions in globally averaged basalt and andesite, based on TES spectra deconvolved using the procedure of Wyatt et al. (2001), are An$_{53}$ and An$_{66}$, respectively.

The Mars Pathfinder alpha proton x-ray spectrometer (APXS) measured the abundances of major elements in rocks on the martian surface, by monitoring their response to alpha particles generated by radioactive decay. Characteristic x-rays are most useful for heavier elements and scattered alpha particles provide data on lighter elements; however, only the surfaces of the rocks are analyzed. The rocks analyzed by APXS are partly coated with sulfur-rich dust, but extrapolation to zero sulfur allows an estimate of the composition of the
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Tharsis and Elysium are thickly blanketed with dust (Simpson et al., 1992; Christensen and Moore, 1992), so the compositions of their volcanic units cannot be determined from spectra. The volcanoes resemble terrestrial basaltic shield volcanoes, but estimates of flow rheology permit magma compositions with a wide range of silica contents (e.g., Zimbelman, 1985; Cattermole, 1987). The huge diameters of summit calderas (Wilson et al., 2001) and analysis of stress patterns around the calderas (Zuber and Mougins-Mark, 1992) on Tharsis volcanoes suggest that magma chambers are significantly larger than on Earth. Numerous episodes of magma intrusion into these reservoirs are implied by overlapping calderas and by models for the construction of these volcanic edifices (Wilson et al., 2001).

![Diagram](image_url)

**Fig. 4.** Mg/Si vs. Al/Si in Mars surface materials and martian meteorites. Depletion in aluminum, relative to terrestrial rocks, is a defining characteristic. Lithologic symbols as in Fig. 1. After Rieder et al. (1997), with revised Mars Pathfinder data from Wänke et al. (2001) and additional meteorite analyses from Warren and Kallemeyn (1997), Dreibus et al. (2000), Taylor et al. (2000), and Zipfel et al. (2000).
Composition of the Crust from Martian Meteorites

Orthopyroxenite (ALH 84001) cannot be petrologically representative of the ancient crust, but formation of this meteorite by crystal accumulation in basaltic magma (Mittlefehl dt, 1994) is consistent with the MGS TES identification of basalt in the southern highlands. Martian volcanic terrains having thermal emission spectra like shergottites, nakhlites, or Chassigny (Hamilton et al., 1997) have not been found (Hamilton et al., 2001b), at least at TES spatial resolution (several kilometers). By default, this leaves dust-obscured (and geologically young) Tharsis as the likely source for all SNC meteorites but ALH 84001.

All martian meteorites have modal pyroxene > plagioclase and correspondingly low Al contents (Fig. 4). This contrasts with plagioclase > pyroxene in TES-derived modes for martian basalt and andesite, reinforcing the idea that SNC meteorites are from Tharsis. Nevertheless, the sodic plagioclase compositions in SNC meteorites (An45-65 for shergottites and ALH 84001, ~An25 for nakhlites; McSween and Treiman, 1998) are similar to those estimated from deconvolved TES spectra. Agreement between remote sensing and laboratory measurements of this important crustal component indicates that Mars is not depleted in volatiles.

Longhi (1991) suggested that mirror-image trace element patterns for nakhlites and basaltic shergottites indicate nakhlite magmas formed by partial melting of the same mantle source that later produced shergottite magmas, consistent with their suggested derivation within the Tharsis plume. However, new data call this hypothesis into question. Figure 5 compares the measured ratios Zr/Hf and Nb/Y, which should be unaffected by igneous fractionation. Distinct ratios for the nakhlites and the shergottites least contaminated by crust (QUE 94201 and DaG 476, see below) suggest different mantle sources. Differences in initial Os isotopic compositions (Brandon et al., 2000) also support the idea of distinct mantle reservoirs for shergottites and nakhlites. Derivation of shergottites and nakhlites by melting within the long-lived Tharsis plume apparently requires sampling of multiple source regions, probably at different depths.

Magmatic Processes that Affected SNC Meteorites

The Iherzolitic shergottites are cumulate plutonic rocks (McSween and Treiman, 1998) and erupting basaltic shergottites carried cumulus crystals that formed at depth (McCoy et al., 1992; Lentz and McSween, 2000), demonstrating that these magmas were modified by fractional crystallization. The absence of olivine and the high incompatible element concentrations in most basaltic shergottites also indicate that these magmas were fractionated. Spacecraft imagery suggests large, complex magma chambers beneath Tharsis volcanoes, allowing ample opportunity for subsurface magmatic processes like fractional crystallization.

The restricted compositions of erupted terrestrial basalts have been attributed to differences in the ability of fractionated magmas to ascend through the crust (Stolper and Walker, 1980). The limited range of molecular Fe/(Mg + Fe) in mid-ocean ridge basalt (MORB) magmas, the Earth’s most common lavas, is illustrated in Fig. 6. During fractionation, the densities of MORB liquids progressively decrease and then increase (Fig. 6), and the compositions of erupted MORB correspond to the minimum density (~2.75 g/cm³). The Earth’s crust apparently acts as a density filter, resulting in a window of eruptibility. Fe/(Mg + Fe) values and calculated densities for shergottite magmas are also illustrated in Fig. 6, along with trajectories for their liquid lines of descent (calculated using the MELTS program). Martian magmas tend to be denser than their terrestrial counterparts because of their high iron concentrations. Like terrestrial basalts, these erupted martian magmas exhibit a restricted density range (Fig. 6), although it corresponds to a somewhat greater compositional range. Density limits suggest that the basaltic crust of Mars, at least in the Tharsis region where the crust is especially thick, may also serve as a density filter for ascending magmas. Fractionation and density control of magmas outside of Tharsis might also explain the scarcity of olivine (i.e., the rarity of unfractionated magmas, in TES orbital maps).

The most notable geochemical differences among shergottites can be explained by varying degrees of magma contamination by a crustal component (Jones, 1989; Longhi, 1991; Borg et al., 1997; Blechert-Toft et al., 1999). This component had high initial 87Sr/86Sr, low initial 143Nd/144Nd, and low initial 176Hf/177Hf. Correlations of radiogenic isotopes with redox
state, long time-integrated enrichment of light REE, and possibly oxygen isotopes (Fig. 7) demonstrate that this ancient crustal component was oxidized, enriched in incompatible elements, and perhaps hydrated, relative to mantle-derived magmas. Crustal assimilation by magmas cannot be observed by remote sensing, but it is to be expected in long-lived Tharsis igneous systems. However, it is surprising that the youngest lavas (Shergotty, Zagami, Los Angeles) assimilated more crust than older shergottites. On Earth, the earlier erupting magmas in a given volcanic center often show a greater degree of crustal contamination.

### COMPOSITION AND DYNAMICS OF THE MARTIAN INTERIOR

#### Geologic Context

The accepted value for Mars' mean density, based on spacecraft measurement of its mass, is $3.9335 \pm 0.0004 \text{ g/cm}^3$. The low mean density can be accounted for if Mars has a lower bulk Fe content than Earth or is enriched in a light element like O (i.e., is more highly oxidized). The planet's dimensionless moment of inertia has been calculated from an improved estimate of the spin pole precession rate determined from Doppler and range measurements to the Mars Pathfinder lander (Folkner et al., 1997). The derived value ($C/MR^2 = 0.3662 \pm 0.0017$) corresponds to a core radius of approximately 1300–1500 km, depending on core composition.

Bands of alternating magnetic polarity, mapped in the southern hemisphere by MGS (Acuna et al., 1999; Purucker et al., 2000), provide additional information on the nature of the core. Mars presently lacks an internally generated dipole magnetic field, so the anomalies must represent remnant magnetism from an ancient magnetic field. The magnetism must arise within the crust, because MGS images of the magnetized region show no correlation with surface features and the depth of the Curie temperature for demagnetization has been estimated at ~35 km, near the bottom of the crust in this region (Nimmo and Gilmore, 2001).

Models of the thermal evolution of Mars, constructed for a range of initial conditions, mantle geometries and viscosities, and heat production scenarios, indicate declining heat flow with time (Schubert et al., 1992; Grasset and Parmentier, 1998). The critical heat loss mechanism is convection, which would have caused pressure-release melting to produce volcanic plains early in martian history (Weitzman et al., 2001). Subsequent (post 1–2.5 Ga) melting would have resulted from heating within plumes (Solomon and Head, 1982). The existence of a perovskite stability field in the lower mantle affects plume production. Convection cells that contain the spinel-to-perovskite transition will preferentially develop long-wavelength flow (Harder and Christensen, 1996; Breuer et al., 1998) and thus tend to form a single, persistent upwelling which
Fig. 7. Chemical and isotopic correlations among basaltic shergottites. Decreasing values of $\epsilon^{143}$Nd (data sources summarized by Wadhwa, 2001) serve as an index for degree of assimilation of crust. The top box shows a correlation with magmatic redox state, as determined from the size of the Eu anomaly in pyroxene (Wadhwa, 2001). The middle box shows the ratio of light to heavy REE (data from Lodders, 1998; Rubin et al., 2000; Zipfel et al., 2000; Nazarov et al., 2001, unpubl. data), which reflects fractionation of incompatible lithophile elements. The lower box indicates deviation of the bulk-rock oxygen isotopic composition from the terrestrial mass fractionation line (data from Clayton and Mayeda, 1996; Rubin et al., 2000; Nazarov et al., 2001, unpubl. data shown as filled symbols; data from Franchi et al., 1999 and Wiechert et al., 2001 shown as open symbols). The open-symbol data have been used to argue that no deviations exist (the shaded horizontal bar contains all SNC analyses by Franchi et al., 1999), so the latter effect is questionable. These correlations (and others, such as initial $87Sr/86Sr$) suggest assimilation, by a mantle-derived magma like QUE 94201, of a crustal component that was oxidized, enriched in incompatible elements, and possibly hydrated with isotopically heavy water.

could explain Tharsis. The crustal dichotomy might also result, in part, from mantle convection. Hemisphere-scale convection is possible if the martian mantle is layered in viscosity, as occurs in the Earth’s mantle (Zuber, 2001).

Sleep (1994) invoked plate tectonics to account for the crustal dichotomy, and Acuna et al. (1999) suggested that magnetic lineations might reflect tectonic spreading. However, Mars lacks the surface manifestations of subduction, and orbital observations do not generally support these speculations.

Mantle and Core Compositions from Martian Meteorites

The compositions of the mantle and core are best estimated using elemental and isotopic abundances in martian meteorites. The canonical model for the martian mantle and core (Dreibus and Wänke, 1985; Wänke and Dreibus, 1988) is based on element ratios in SNC meteorites (Table 1). This model calls for accretional mixing of reduced and oxidized components, the latter containing volatiles in Cl-chondrite proportions. Several studies (Longhi et al., 1992; Sohl and Spohn, 1997) have used this model composition to calculate the interior structure of Mars, and Bertka and Fei (1997) experimentally determined mantle mineral stabilities using this composition.

The experiments (Fig. 8) suggest that the uppermost mantle consists of olivine + clinopyroxene + orthopyroxene + garnet. A transition zone between 14 and 23 GPa is composed mostly of $\gamma$-spinel and majorite. The presence of a lower mantle containing Mg-Fe silicate perovskite is sensitive to the temperature profile, with stishovite substituting for perovskite at lower temperatures. Perovskite is more likely to have occurred early in Mars history, when mantle temperatures were hotter. Regardless of the temperature profile, the lower mantle accounts for a smaller proportion of the interior of Mars than does the corresponding region within the Earth.

Two other compositional models for Mars are based on matching its oxygen isotopic composition, as determined from
TABLE 1. Estimated mantle + crust and core compositions (wt%), density, and moment of inertia.

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<td>n.d.</td>
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<tr>
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<tr>
<td>Olivine</td>
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<tr>
<td>Garnet</td>
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<tr>
<td>Other†</td>
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<td>1.6</td>
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<td>C/MR²</td>
<td>0.367*</td>
<td>0.367</td>
<td>0.361</td>
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*Models are WD (Dreibus and Wänke, 1985; Wänke and Dreibus, 1988), LF (Lodders and Fegley, 1997), and S (Sanloup et al., 1999).
†Normative mineralogy at 3 GPa calculated using the algorithm of McGechin and Smyth (1978).
‡Other normative minerals include ilmenite, chromite, and whitlockite.
§Mantle densities calculated using mineral densities from Robie and Hemingway (1995).
#Core densities calculated using mineral densities from Lodders and Fegley (1997).
*This value differs from that (0.353) calculated by Longhi et al. (1992) using the WD model, presumably because of choice of definitions for normative compositions and mineral densities. Moment of inertia is also sensitive to the assumed crust thickness.

SNC meteorites (Clayton and Mayeda, 1996), by mixing various chondrite classes. (Note that in the Wänke-Dreibus model, the reduced component is not identified with a specific chondrite type and thus the oxygen isotopic composition of a mixture could conceivably match SNC meteorites.) The model of Lodders and Fegley (1997) mixes 85% H, 11% CV, and 4%

CI chondrites to obtain the requisite oxygen isotopic ratios, with the bulk composition (Table 1) following from mass balance calculations. The model preferred by Sanloup et al. (1999) (Table 1) combines 55% H and 45% EH chondrites in the same manner. All Mars compositional models share the characteristic that the martian mantle is more Fe-rich than in the Earth, consistent with the Fe-rich compositions of SNC meteorites. However, the high-pressure normative mineralogies of the silicate portion (mantle + crust) for these models vary (Table 1).

Treiman et al. (1987) estimated that the core constituted ~30% of the mass of Mars, based on partitioning of siderophile elements in SNC meteorites. A somewhat smaller core (Table 1) was suggested by Wänke and Dreibus (1988) based on depletion of chalcophile elements in SNC meteorites. Both these estimated core compositions are S-rich. The chondrite mixing models of Lodders and Fegley (1997) and Sanloup et al. (1999) calculated S directly from mass balance; their models differ significantly in core mass and S abundance (Table 1). The S content of the core has implications for its physical state (solid or liquid); the present-day core would likely be liquid if its S content were ≥15% (Schubert et al., 1992). Sulfur abundance also affects core density (Table 1). Measured abundances of siderophile elements in SNC meteorites are
consistent with S-bearing metal-silicate equilibrium at high temperature and pressure. An equilibrium model implies homogeneous accretion (Treiman et al., 1987; Wänke and Dreibus, 1988; Righter and Drake, 1996; Righter et al., 1998) and is consistent with a high degree of melting (i.e., a magma ocean). Warren et al. (1999) measured high abundances of siderophile elements in SNC meteorites and concluded that an accretionary veneer of condritic material was necessary, but other data (Righter et al., 1998; Brandon et al., 2000) may rule out a late veneer.

Table 1 compares calculated mean densities and moments of inertia for the three Mars models. The compressed mean density and moment of inertia for the Sanloup model apparently lie outside accepted values for Mars, whereas calculated values for the other two models are in reasonable agreement with these values. The physical properties of the Wänke–Dreibus model are best constrained, because this composition has been studied experimentally. A derived density profile for Mars, consistent with a range of model core compositions (Bertka and Fei, 1998), is shown in Fig. 8. Using the experimentally determined abundances of mantle minerals and the estimated core mass of this model, Bertka and Fei (1998) calculated a moment of inertia (0.354) that is consistent with the Mars Pathfinder measurement. However, that model requires an unrealistically thick crust (180–320 km). For this reason, Bertka and Fei (1998) argued that a basic feature of the Wänke–Dreibus model, its bulk CI Fe/Si ratio, could not be correct.

Mantle and Core Evolution and Dynamics from Martian Meteorites

Because Hf and W are strongly fractionated into silicate and metal, respectively, the $^{182}\text{Hf}^{182}\text{W}$ isotopic system can determine the timing of core formation. Isotopic data for SNC meteorites (Lee and Halliday, 1997) indicate early core separation within 30 Ma of solar system formation. This chronologic constraint on core formation is consistent with $^{148}\text{Sm}^{142}\text{Nd}$ data from the same meteorites (Harper et al., 1995; Borg et al., 1997), which reflect early depletion of light REE during formation of the crust. The initial $^{187}\text{Os}^{188}\text{Os}$ ratios for SNC meteorites (Brandon et al., 2000) also vary systematically with $^{182}\text{W}$ and $^{142}\text{Nd}$ (Fig. 9). Synchronous differentiation of core and crust, as suggested by these correlations between Nd, W, and Os isotopic anomalies, would have been accomplished most easily in a magma ocean. The isotopic evidence for early planetary differentiation is consistent with spacecraft observations of widespread Noachian crust and an early magnetic field, but there are no direct observations supporting a magma ocean (Zuber, 2001).

The preservation of Nd and W isotopic anomalies from the decay of extinct radionuclides, plus correlating isotopic anomalies in the long-lived Re/Os system (Brandon et al., 2000), indicate that large-scale, vigorous mantle convection and crustal recycling have not occurred or were very short-lived. The long-term depletion of incompatible elements in the SNC mantle source region (Longhi, 1991) and the ancient isotope model ages for SNC meteorites (Chen and Wasserburg, 1986; Borg et al., 1997; Blichert-Toft et al., 1999) strongly support this conclusion. SNC meteorites are apparently derived from mantle sources that differentiated early and then remained largely inert for most of the planet's history. The limitation on isotopic mantle convection seems inconsistent with the suggestion that most martian meteorites formed in the convective Tharsis plume.

Oxides and sulfides within the carbonates of ALH 84001 carry the rock's palaeomagnetic record. Weiss et al. (2001) argued that the magnetization originated during a shock-melting event which is constrained to ~4 Ga by Ar isotopic data. Given
the intensity of the measured magnetic moment, this sample was probably magnetized by a geodynamo or by nearby crustal remanent fields like those observed by MGS in the southern hemisphere. Since the crustal fields were also created by a dynamo, this provides strong support for an early martian magnetic field, and presumably an early convecting core that was at least partly molten.

The magnetic anomalies detected by MGS have been modeled as arising from 30 km thick crustal slabs having high remanent magnetic strength (Acuna et al., 1999). The magnetic minerals in younger SNC meteorites (titanomagnetite and pyrrhotite) do not have the compositional purity and single-domain particle sizes required to account for this magnetism, nor is this natural remanent magnetism stable enough to survive for billions of years (Hargraves et al., 2001). Magnetites in ALH 84001 are relatively pure and have single-domain particle sizes (Bradley et al., 1996; Thomas-Keptka et al., 2000). However, disputes about how these magnetites formed cloud whatever relevance they might have to paleomagnetism in the deep martian crust.

MARS DEGASSING, HYDROGEOLOGY, AND WEATHERING

Orbital Geomorphology

SNC Volatiles Water

Soil Stable Isotopes Weathering

Geologic Context

Viking Orbiter and MGS images of volcanic deposits interpreted as pyroclastic materials (Mouginis-Mark et al., 1992b) demonstrate that planetary degassing has occurred. As a consequence of the lower martian gravity and the fact that volatile solubility in magmas is dependent on pressure, gas exsolution from ascending martian magmas would have begun at greater depths than on the Earth and volatile release would have been more efficient (Head and Wilson, 1989). The amount of outgassed water has been estimated from a variety of spacecraft observations and measurements. (This quantity is commonly expressed in terms of a globally distributed water layer of uniform depth; for comparison, the Earth has outgassed 2700 m of water.) Constraints from the isotopic compositions of atmospheric gases (Donahue, 1995) and extrapolated from the volume of eroded materials (Carr, 1987) bracket the amount of water between a few tens of meters to hundreds of meters. Intermediate values 120–150 m are based on the release of magmatic gases from magmas having water contents comparable to terrestrial lavas (Greeley and Schneid, 1991; Phillips et al., 2001).

Evidence that rocks on the martian surface interacted with water is provided by orbiter imagery. Noachian terrains in the southern highlands are incised by networks of tributaries, and large buried channels once emptied into the northern lowlands (Baker, 2001). Most of the erosion features are concentrated within the trough encircling Tharsis, suggesting a connection between volcanism and the release of water (Phillips et al., 2001). Flooding of the lowlands may have produced an ancient ocean, now represented by a basin with a very smooth floor which is bounded by possible shoreline features (Parker et al., 1993) that approximate an equipotential surface (Head et al., 1999). Other watersheds drained into Hellas and Argyre (Smith et al., 1999), and evidence for past lakes abounds in smaller impact craters (Baker, 2001). MGS imagery shows widespread layering that has been interpreted as ancient, water-lain sediments (Malin and Edgett, 2000a). Now, however, most martian water must be underground. Patterned ground and rampart craters are consistent with permafrost (Baker, 2001), and the existence of groundwater or subsurface ice is supported by the MGS observation of modern seepages from the walls of craters and valleys (Malin and Edgett, 2000b).

The planetary CO$_2$ inventory has been estimated from C/N ratios of possible accreted volatile sources (comets or carbonaceous chondrites). CO$_2$ can be removed from the atmosphere by dissolution in water and precipitation of carbonates. These CO$_2$ estimates, expressed as equivalent global thicknesses of carbonate, vary from 3 to 20 m (Bogard et al., 2001). However, carbonate has not been detected spectroscopically, and C in soils at the Mars Pathfinder site is <0.3 wt% (Foley et al., 2000). Similarly, other evaporite minerals have not been identified in orbital thermal emission spectra, but soils at the Viking and Mars Pathfinder sites have abundant S and Cl, interpreted to occur as sulfates and chlorides (Clark et al., 1982; Rieder et al., 1997).

The pervasive, fine-grained dust that blankets much of the planet's surface has not been characterized mineralogically, although it is commonly thought to consist of the weathering products of volcanic rocks. Similarity in the measured chemical composition of soils from widely separated landing sites (Bell et al., 2000; McSween and Keil, 2000; Wänke et al., 2001) suggests that this globally distributed unit has been homogenized by eolian activity. Differences in multispectral properties among chemically similar soils at the Mars Pathfinder site are probably related to variations in redox levels of trace pigimentary minerals or physical properties, rather than mineralogy (Bell et al., 2000). Absorption features in lander surface spectra (Bell et al., 2000) and magnetics experiments (Madsen et al., 1999) indicate the presence of nanophase ferric oxides which suggest chemical weathering. Deposits of crystalline gray hematite recognized from MGS TES measurements in Sinus Meridani (Christensen et al., 2000b) and dispersed ferricydrate in soils identified by orbital VISNIR
spectroscopy (Murchie et al., 2000) are also attributed to aqueous processes. The response of Viking fines to heating (Biemann et al., 1977) and simulations of surface reflectance spectra suggest soil water contents of a few percent (Yen et al., 1998).

**Volatile Inventory and Degassing from Martian Meteorites**

The Wänke–Dreibus model supposes that Mars originally accreted with a significant volatile component. However, based on the abundance of CI in SNC meteorites and relative solubilities of CI and H₂O in basaltic magma, this model predicts a dry mantle containing only 36 ppm water. Lodders and Fegley (1997) argued that an analogous calculation would give an incorrect estimate of water in the Earth's mantle, because CI has been sequestered in the crust. The Lodders–Fegley model suggests a bulk Mars CI abundance 8x higher, with a correspondingly higher mantle water content.

Oxygen and hydrogen isotope systematics in SNC meteorites suggest a significant reservoir of crustal water (Karlsson et al., 1992; Leshin et al., 1996), as indicated by spacecraft observations. The volume of water in the crust has been estimated from a model attributing the atmospheric D/H ratio to preferential loss of the light isotope from water having D/H as measured in martian meteorites. Leshin (2000) suggested a crustal reservoir twice as large (equivalent to 50–200 m) as previously thought based on atmospheric D/H (Donahue, 1995).

The low measured bulk H₂O contents of SNC meteorites have been cited as evidence that their parent magmas and the source region that melted to produce them were dry. However, several petrologic observations suggest that these magmas contained appreciable water prior to eruption. Melt inclusions trapped within SNC phenocrysts that crystallized at depth sometimes contain amphiboles. Attempts to model the solidification of amphibole-bearing melt inclusions and thereby to constrain magma water contents at the time of trapping (Johnson et al., 1991; McSween and Harvey, 1993) give estimates ranging from 1 to 2 wt% H₂O. However, conflicting interpretations of the low measured H contents of the amphiboles (Watson et al., 1994; Mysen et al., 1998; King et al., 1999) may render this evidence inconclusive. Measurements of soluble trace elements in Shergotty pyroxene phenocrysts suggest a higher dissolved water content for its magma at depth (McSween et al., 2001), and experiments that crystallize Shergotty pyroxene compositions imply a pre-eruptive H₂O content of ~1.8 wt% (Dann et al., 2001). This water was presumably lost during magma ascent and eruption (i.e., it was delivered to the planet's surface). Fractionation of hydrous basaltic magma would provide a way to account for large volume of andesitic magma that might have erupted on Mars, because hydrous magmas require less fractionation to reach andesite composition (Minitti and Rutherford, 2000; Dann et al., 2001). Water in the Shergotty magma could have been derived either from the mantle or by assimilation of hydrous crustal materials. The possibility of hydrous shergottite magmas provides a way to reconcile geologic and geochemical constraints on martian degassing.

SNC meteorites provide precise estimates of N and noble gas atmospheric abundances, as well as information on mantle gas components. Atmospheric ⁴⁰Ar represents <2% of that produced by radioactive decay of ⁴⁰K during the last 4.0 Ga (Bogard et al., 2001). This comparison suggests either very limited planetary outgassing or efficient loss of atmospheric gases by sputtering and impact erosion. Retention of a significant concentration of primordial ³⁶Ar in the mantle component of SNC meteorites (Bogard et al., 2001) is consistent with a relatively undegassed interior, whereas atmospheric loss is favored by the high ¹⁵N/¹⁴N (Marti and Mathew, 2000) and ¹²⁹Xe/¹³²Xe (Musselwhite et al., 1991) ratios of trapped gases in SNC meteorites. It is difficult to reconcile limited planetary outgassing with extensive melting to form a voluminous crust.

**Fluid–Rock Interactions in Martian Meteorites**

Carbonates in ALH 84001 were deposited in fractures at ~4 Ga (Borg et al., 1999), probably either by evaporation of brines (McSween and Harvey, 1998; Warren, 1998) or reaction with hydrothermal fluids (Romanek et al., 1994). Nakhlites contain "iddingsite" consisting of smectite, ferrihydrite, and magnetite (Treiman et al., 1993), as well as salts, including sulfates, carbonates, and halite (Bridges and Grady, 2000). The formation of iddingsite in Nakhla has been radiometrically dated at ~650 Ma (Swindle et al., 2000). Aqueous fluids also percolated through shergottites, depositing small amounts of clays, calcite, sulfates, phosphate, and halite (e.g., Gooding et al., 1988). These secondary minerals confirm the existence of saline groundwater inferred from spacecraft imagery and are consistent with the oxidized nature and high abundance of S and CI in martian soils (Bell et al., 2000; McSween and Keil, 2000).

Isotopically heavy O (Karlsson et al., 1992; Romanek et al., 1998) and H (Leshin, 2000) in SNC meteorites suggest that water has been recycled from the atmosphere, where light isotopes were depleted, through the hydrosphere and ultimately back into the lithosphere. Differences in the S isotopic composition of sulfides in nakhlites and Chassigny have been explained by the interaction of rock with hydrothermal fluids (Greenwood et al., 2000). Mass-independent isotopic fractionation of S in SNC meteorites can only be accounted for by reactions in the atmosphere (Farquhar et al., 2000), so much too must have been recycled. The existence of volatile cycles is not apparent from remote sensing, but can be inferred from evaporation of fluids in seepages (Malin and Edgett, 2000b) and observations suggesting that surface waters have recharged the groundwater system (Heap and Pratt, 2001). Recycling of atmospheric volatiles into the lithosphere must occur by mechanisms other than subduction.
SUMMARY

The rocks of Mars have been studied from afar by spacecraft and by close inspection in the laboratory. For the most part, first-order conclusions drawn from both kinds of observations and measurements are in agreement. The crust covering most of the planet is ancient and thick. The chemical compositions of the mantle and core are reasonably well constrained, and their derived mineralogies are consistent with geophysical constraints. Mantle homogenization through convection and crustal recycling has not occurred. Our understanding of the volatile inventory and degassing history of the planet, as well as its hydrology and surface weathering processes, are reinforced by conformity in these data sets.

However, some substantial differences arise in interpreting crust compositions and crystallization ages, because martian meteorites constitute a biased and unrepresentative sampling of the lithosphere. All SNC meteorites but one were probably derived from Tharsis or Elysium, and oversampling of this terrain suggests that the ejection mechanism does not usually deliver samples of the pervasive Noachian crust. Because late plume volcanism is likely to have differed from the early igneous and sedimentary processes that shaped most of the planet's surface, conclusions about crustal petrogenesis drawn from these meteorites are likely to be incomplete.

Determining if martian meteorites conform to the expectations of martian geology provides a test of whether the puzzle is being assembled correctly. Viewed from this complementary perspective, the emerging Mars puzzle is progressing well.

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REFERENCES


