1. DISCOVERY AND EXPLORATION

1.1. Discovery and Early Groundbased Observations

Once the Galilean satellites of Jupiter were discovered and the Copernican model of the solar system became widely accepted, Mars’ apparent lack of a moon was notable. By the end of the seventeenth century, Saturn was known to have at least five satellites, a number that swelled to seven by the end of the eighteenth century. By 1800 a new planet, Uranus, was discovered and found to have at least two satellites. By 1870, two more uranian satellites and another saturnian satellite were detected, along with another new planet (Neptune) with its own satellite. Yet Mars remained moonless. Based on numerology, Kepler predicted that Mars should have two satellites, as two made the most sense when interpolating between Earth’s single moon and Jupiter’s four. Jonathan Swift and Voltaire both “predicted” that Mars would have two moons, but their predictions were based on satire (Hall, 1878).

Both William Herschel and Heinrich Louis d’Arrest performed unsuccessful searches for martian moons before Asaph Hall, using the U.S. Naval Observatory 26-inch (66-cm) refractor, found first Deimos then Phobos during the excellent Mars apparition of 1877. Because of the primitive state of astrophotography, only visual observations using eyepieces were possible at the time. The satellites were named for characters in the Iliad, Phobos (Fear) and Deimos (Terror), who are the attendants of the god Ares, the Greek equivalent of the Roman god Mars. Despite searches for additional satellites, using both ground- and spacecraft-based observations, no additional satellites of Mars have been found within its Hill sphere to a diameter of 180 m assuming an albedo of 0.07 (Sheppard et al., 2004).

Given the close proximity of Phobos and Deimos to Mars, positional measurements were typically the only ones undertaken. Pascu et al. (2013) compiled and reviewed these measurements as well as other early observations. A focus of early observations was the secular acceleration of Phobos in its orbit (e.g., Sharpless, 1945), now known to originate from tidal effects that cause the moon to spiral in toward Mars, which eventually will result in Phobos’ impact (Burns, 1978). The first photometric study to estimate the moons’ diameters from their brightnesses, by Edward Charles Pickering of the Harvard College Observatory and reported by Hall (1878), suggested diameters of 9 km for Deimos and 11 km for Phobos. These assumed a Mars-like albedo and hence resulted in diameters smaller than the actual values. Just prior to the first spacecraft encounters with the moons, a combination of photoelectric photometry (Harris, 1961) and albedo estimates from polarimetry (Zellner, 1972) resulted in diameter estimates within 10% of current values (Zellner and Capen, 1974).
1.2. 1969–1980: Exploration by the Mariner and Viking Spacecraft

It was not until the visit of spacecraft to Mars that bulk properties and the physical geology of the moons would become known (Table 1, Figs. 1 and 2). The first image of Phobos, by Mariner 7, was made in silhouette against the martian surface. *Smith* (1970) found an albedo of 0.065 from those data, at that time the lowest-known albedo in the solar system. The first images resolving surface features, from Mariner 9, established approximate dimensions of the two moons, their heavily cratered surfaces, low albedos, and coverage by regolith (*Pollack et al.*, 1972, 1973). Polarimetric measurements indicated a finely granular regolith (*Noland et al.*, 1973).

Higher-resolution imaging from close flybys of the moons by the Viking 1 and Viking 2 orbiters revolutionized understanding of both moons (*Duxbury and Veverka*, 1977), and still provides most of the current knowledge about the physical geology of Deimos. Results through Viking were thoroughly reviewed by *Thomas et al.* (1992) and *Burns* (1992). Phobos’ surface exhibits rough topography, including craters in a variety of degradation states. One moderately well preserved large crater, Stickney, is about 9.4 km in diameter. Reflectance varies spatially, with brighter materials apparently less affected by “space weathering” being exposed on crater rims (*Thomas*, 1979). [Space weathering refers to processes that alter the optical properties of regolith over geologic time. On the Moon, this includes darkening accompanied by reddening of the spectrum and weakening of mineral absorptions. It is understood to occur by formation of nanophase metallic iron by redeposition of vapor from micrometeoroid impacts (*Hapke*, 2001).] Topographic grooves 100–200 m wide, tens of meters deep, and kilometers in length form several subparallel sets, most of which define planes that are nearly parallel to Phobos’ intermediate axis (i.e., to the direction of orbital motion) (*Thomas et al.*, 1979). In contrast, Deimos has a markedly smoother surface (Fig. 3) with craters that are more degraded and infilled by regolith (*Thomas*, 1979). Migration of regolith down slopes has formed elongate streamers (*Thomas and Veverka*, 1980; *Thomas et al.*, 1996). Grooves have not been recognized on Deimos.

Several spectra of Phobos were acquired at ultraviolet through near-infrared wavelengths by Mariner 9 and Viking, and assembled into a single, composite spectrum

![Fig. 1. Phobos image mosaic (Stooke, 2011) projected on a shape model from Gaskell (2011). Upper left: Southern hemisphere. Upper right: Northern hemisphere. Center left: Leading hemisphere. Center right: Sub-Mars hemisphere. Lower left: Trailing hemisphere. Lower right: Anti-Mars hemisphere.](image1)

![Fig. 2. Deimos image mosaic projected on shape model, both from Thomas (1993). Upper left: Southern hemisphere. Upper right: Northern hemisphere. Center left: Leading hemisphere. Center right: Sub-Mars hemisphere. Lower left: Trailing hemisphere. Lower right: Anti-Mars hemisphere.](image2)
by Pang et al. (1978). That composite spectrum is flat at wavelengths longer than 0.4 µm but falls off steeply into the ultraviolet. Similarity to the reflectance spectrum of carbonaceous chondrite meteorites drove a long-standing hypothesis that the moons are primitive, carbonaceous bodies captured into orbit around Mars (Pang et al., 1978; Pollack et al., 1978). Alternatively, it has been proposed that the moons accreted in Mars orbit from material like that forming bulk Mars (Burns, 1992), or from ejecta from a large impact with Mars (Singer, 1966; Craddock, 2011), and that they are dark because of extreme space weathering (Britt and Pieters, 1988).

### Measurement of the moons' masses, made by tracking perturbations of the Viking Orbiters’ trajectories during close flybys, combined with volume estimates from imaging, yielded the first density estimates of 2000 ± 500 kg m$^{-3}$ for Phobos and 1900 ± 700 kg m$^{-3}$ for Deimos (Veverka and Burns, 1980). Although subsequent data and new analyses improved these values (Table 1), the Viking results clearly showed that Mars’ moons are much less dense than the bulk rock making up Mars, requiring that they must be either highly porous, composed of lower-density material, or both.

### 1.3. 1980–2004: Incremental New Understanding

The next phase of exploration included groundbased observations that refined earlier understanding of the moons’ properties, a partially failed close rendezvous, and distant observations by other Mars-orbiting or landed spacecraft. Earth-based spectra revealed that the 3-µm H$_2$O absorption expected for a hydrated-mineral-bearing carbonaceous composition was undetectable (Bell et al., 1989; Rivkin et al., 2011).

### Table 1. Current best estimates of global properties of Phobos and Deimos.

<table>
<thead>
<tr>
<th>Property</th>
<th>Phobos</th>
<th>Deimos</th>
</tr>
</thead>
<tbody>
<tr>
<td>Orbital semimajor axis</td>
<td>9377 km (2.8 R$_{\text{Mars}}$)</td>
<td>23,460 km (7 R$_{\text{Mars}}$)</td>
</tr>
<tr>
<td>Orbital period</td>
<td>7.66 h</td>
<td>30.3 h</td>
</tr>
<tr>
<td>Orbital eccentricity</td>
<td>0.0151</td>
<td>0.0003</td>
</tr>
<tr>
<td>Orbital inclination, to Mars’ equator</td>
<td>1.093°</td>
<td>0.93°</td>
</tr>
<tr>
<td>Rotational period</td>
<td>Synchronous</td>
<td>Synchronous</td>
</tr>
<tr>
<td>Size</td>
<td>$26.06 \times 22.80 \times 18.28$ km</td>
<td>$15.0 \times 12.2 \times 10.4$ km</td>
</tr>
<tr>
<td>Density</td>
<td>$1860 \pm 13$ kg m$^{-3}$</td>
<td>$1490 \pm 190$ kg m$^{-3}$</td>
</tr>
<tr>
<td>Gravity</td>
<td>$3 \times 10^{-3}$ m s$^{-2}$</td>
<td>$2 \times 10^{-3}$ m s$^{-2}$</td>
</tr>
<tr>
<td>Normal reflectance, 0.55 µm</td>
<td>$0.071 \pm 0.012$</td>
<td>$0.068 \pm 0.007$</td>
</tr>
</tbody>
</table>

* Wilner et al. (2014).
† Thomas (1993).
§ Simonelli et al. (1998).
¶ Thomas et al. (1996).

**Fig. 3.** See Plate 16 for color version. Spacecraft exploration reveals global properties of Phobos and Deimos, and basic differences between them. **Top left:** Viking images revealed that Phobos’ surface is relatively rough, with well-preserved craters, and parallel topographic grooves. **Top center:** Deimos’ surface is smooth with craters infilled by regolith. **Top right:** Multispectral and hyperspectral imaging from Phobos 2, Mars Express, and the Mars Reconnaissance Orbiter (shown, centered on the sub-Mars hemispheres) reveal spatial variations in Phobos’ spectral properties, especially associated with Stickney. Deimos’ brightness variations are accompanied by less color variation. **Middle:** Phobos simple cylindrical image map and dynamical height, using shape model from Gaskell (2011) and image mosaic from Stooke (2011). Red is high, blue low, with a range of heights of 1.8 km. **Bottom:** Deimos simple cylindrical image map and dynamical heights, using shape model and image mosaic updated from Thomas (1993). Red is high, blue low, with a range of heights of 1.9 km.
2002), and that disk-integrated spectra of Deimos (Grundy and Fink, 1992) and Phobos (Murchie and Zellner, 1994) were featureless and much redder than previously thought. In hindsight, the shape of the composite spectrum of Pang et al. (1978) was a result of mixing shorter wavelengths that covered a more red-sloped part of Phobos and longer wavelengths covering a much less red part of Phobos (Murchie and Erard, 1996).

The Phobos 2 mission (Sagdeev and Zakharov, 1989) was to have conducted proximity measurements of Phobos’ chemical composition, but the spacecraft was lost after its first month of distant observations, before closest approach to the moon. Three medium-distance encounters with Phobos that did occur provided the most precise estimate of density to date (Avanesov et al., 1991). During those encounters, color imaging returned by the Videospectrometric Camera (VSK) (Avanesov et al., 1991) and ultraviolet through near-infrared spectra by the Combined Radiometer and Photometer for Mars (KRFM) (Ksanfomality et al., 1990) and Imaging Spectrometer for Mars (ISM) (Bibring et al., 1989) provided the first spatially resolved measurements of spectral properties of a low-reflectance body. Phobos’ surface was revealed to be heterogeneous, with a “redder unit” occupying most of the surface, and a “bluer unit” associated with Stickney (Fig. 3). The two units differ mainly in spectral slope, and any mineral absorptions are weak. The 3-µm absorption remained undetected even with improved spatial resolution (Murchie et al., 1991; Murchie and Erard, 1996). The red spectral slopes of the two bodies, plus weak to absent 3-µm absorptions, are more consistent with D-type asteroids than with carbonaceous chondrites (Rivkin et al., 2002).

Measurements by Mars Global Surveyor and Mars Pathfinder added to the growing understanding of the moons. Images from the Mars Orbiter Camera, comparable to the best from Viking, revealed a high spatial density of blocks just east of Stickney, where ejecta from that crater are predicted to be thickest (Thomas et al., 2000). Thermal emission spectrometer (TES) measurements revealed a nearly blackbody spectrum from 10 to 30 µm, possibly with weak superimposed silicate features (Roush and Hogan, 2000, 2001; Gloitch et al., 2015). Whole-disk spectra of both moons from the Imager for Mars Pathfinder (Murchie et al., 1999) confirmed the red-sloped spectra of both moons, and identified a 0.7-µm absorption that they suggested could be due to Fe-bearing minerals.

1.4. Beginning in 2004: Exploration by Mars Express and Mars Reconnaissance Orbiter

Knowledge of Phobos and Deimos was revolutionized again by the Mars Express and Mars Reconnaissance Orbiter (MRO) missions. During multiple close flybys of Phobos, Mars Express (Witasse et al., 2013) acquired nearly global coverage with the High Resolution Stereo Camera (HRSC) (Neukum et al., 2004; Jaumann et al., 2007) and the Observatoire pour la Mineralogie, l’Eau, les Glaces et l’Activité (OMEGA) (Bibring et al., 2004) imaging spectrometer, at multiple viewing geometries that support derivation of the photometric properties of the surface over a broad wavelength range (Fraeman et al., 2012). Additional spectral information came from the thermal infrared Point Fourier Spectrometer (PFS) (Giuranna et al., 2011). The MaRS radio science experiment further improved the accuracy of mass determination, which, combined with a refined global volume estimate from HRSC, yields a density estimate for Phobos with a precision of <1% (Table 1). From low Mars orbit, MRO conducted several observing sessions dedicated to the moons. By virtue of its very high spatial resolution, the High-Resolution Imaging Science Experiment (HiRISE) (McEwen et al., 2007) obtained the highest-resolution color imaging of the moons to date in three broadbands from 0.4 to 1.0 µm. The Compact Reconnaissance Imaging Spectrometer for Mars (CRISM) (Murchie et al., 2007) collected hyperspectral imaging of Phobos complementary to the OMEGA data, and the first spatially resolved hyperspectral measurements of Deimos.

Results from Mars Express and MRO frame the current debates about the composition, geological processes, and origin of the moons, as described in the following sections. In particular, the complexity of the grooves on Phobos frustrates simple explanation of their origins (Ramsley and Head, 2013b; Murray and Heggie, 2014). Similarly, spectral features are weak and have been interpreted as resulting from dissimilar compositions corresponding to the range of hypotheses for the moons’ origins (Giuranna et al., 2011; Fraeman et al., 2012, 2014).

1.5. Relationship to Themes of Solar System Exploration

The origin and evolution of the martian moons provides insight into fundamental questions regarding formation of the terrestrial planets and the geology of asteroidal bodies. For example, questions from Planetary Science Decadal Survey (Squyres et al., 2011) include: (1) What were the initial stages, conditions, and processes of solar system formation? (2) What governed the supply of water to the terrestrial planets? (3) What were their primordial sources of organic matter? (4) What processes have controlled the surface evolution of small bodies?

Table 2 lists five possible origins for Phobos and Deimos, resulting from capture (Pollack et al., 1978) or in situ formation hypotheses (Burns, 1992; Craddock, 2011). Each origin corresponds to a distinct range of compositions, some of which are in the meteorite collection and some of which have only been modeled. If either Phobos or Deimos is a captured carbon- or water-rich body, then that moon may be a sample of the population of bodies that provided carbon and volatiles to the accreting terrestrial planets; such a result would address the second and third Decadal Survey questions. However, capture does not necessarily imply a carbon-rich composition, or even the same origin for Phobos and Deimos. Alternatively, if Phobos or Deimos originated by capture of an ordinary chondrite, that find-
ing would provide new insight into how space weathering affects the link between common meteorites and D-type bodies. Formation by coaccretion with Mars would provide access to a building block of Mars, whereas formation from a large impact would sample Mars’ earliest crust plus the impactor material. Whatever the moons’ origins, addressing this question also addresses the first Decadal Survey question, and characterizing geological processes on Phobos and Deimos addresses the last.

2. PHYSICAL GEOLOGY

The two martian moons differ from one another in appearance, and each has fundamental surface characteristics that remain poorly understood. There was an early expectation that the moons were exemplary of the geology of small bodies in general (Veverka and Burns, 1980), but that expectation has been only partially met, due in part to the moons’ distinct features compared with features observed on asteroids (see the chapters by Yoshikawa et al., Russell et al., and Barucci et al. in this volume). Yet the fundamental influences on all the bodies are similar: impact cratering, orbital mechanics and evolution, low gravity, rotational accelerations, thermal cycling, and solid-body mechanical properties. One of the fundamental questions about Phobos’ and Deimos’ history is, given their similar environments, why does the geology of the surface appear so different?

2.1. Basic Appearance

Phobos is modestly elongated (Table 1), heavily cratered, and distinctively marked with patterns of elongated depressions, termed grooves (Fig. 1). The largest crater, Stickney, is ~9.4 km in diameter, ~0.85× the satellite’s mean radius. This size is within the normal range for the largest craters on small objects (Thomas, 1999; Morrison et al., 2009). The depth of the crater at present is about 1 km, but a large debris accumulation in the center is probably over 400 m thick (Thomas, 1998). Depth-to-diameter ratios of Phobos’ craters appear to be generally 0.2 or less (Thomas et al., 1979; Basilevsky et al., 2014; Kokhanov et al., 2014). The differences between crater morphologies on small objects
and the Moon are still not well established (Morrison et al., 2009; Basilevsky et al., 2014), in part because evaluation of crater degradation states on small objects is not well defined.

Deimos presents a surface whose smoothness contrasts greatly with the cratered and grooved roughness of Phobos (Fig. 2). Albedo features are conspicuous on ridges and some crater rims, further distinguishing its appearance from that of Phobos. It is even more irregularly shaped than Phobos, with a large concavity centered on the south pole (Thomas, 1998).

### 2.2. Shape, Density, and Internal Porosity

The dimensions of Phobos are well known, but Deimos’ are less so. Wilner et al. (2014) used a set of 665 surface control points on Phobos from HRSC and Viking Orbiter images to generate a spherical harmonic function up to degree and order 17 to describe Phobos’ shape. Their best-fit triaxial ellipsoid dimensions (diameters) are $a = 26.06\, \text{km}$, $b = 22.80\, \text{km}$, $c = 18.28\, \text{km}$; the full shape model has a volume of $5742 \pm 35\, \text{km}^3$. These numbers compare closely to the previous dimensions and volume obtained by Duxbury (1989, 1991) and Thomas (1989). Uncertainty in Deimos’ volume and shape result from limited coverage and poor resolution of the trailing hemisphere and southern latitudes (Fig. 2). Its ellipsoidal fit dimensions are reported by Thomas (1993) as $a = 15.6\, \text{km}$, $b = 12.0\, \text{km}$, and $c = 10.2\, \text{km}$ with a volume of $1017 \pm 130\, \text{km}^3$, the latter obtained directly from the shape model.

Phobos’ overall shape is elongated, with the long axis aligned with Mars. Stickney creates a large concavity in the leading part of the sub-Mars hemisphere (Fig. 1). Due to the eccentricity of Phobos’ orbit, the moon librates about its mean position by $\sim 1^\circ$; the exact magnitude of that libration is affected by internal density structure. If Phobos’ interior is homogeneous, then a libration in longitude of $1.10^\circ$ is predicted; however, increased porosity beneath Stickney due to impact fracturing could result in a libration up to $1.24^\circ$ (Rambaux et al., 2012). Previous determinations each have uncertainties of $0.1^\circ$–$0.2^\circ$, and differ from each other by a larger amount than the difference between the two predictions (Duxbury, 1991; Wilner et al., 2014; Oberst et al., 2014). Even the most recent and presumably accurate determination, $1.09^\circ + 0.10^\circ$ (Oberst et al., 2014), is insufficient to distinguish models of interior porosity. In the future, more accurate measurement of libration in longitude may provide a window into heterogeneity in Phobos’ interior structure.

Deimos’ shape can be described as having several rounded facets (Fig. 2). Ridges at the edges of the facets form gravitational highs, and the centers of the facets are gravitationally low (Fig. 3), creating regional slopes (Thomas, 1993; Thomas et al., 1996). High southern latitudes lie in a major concavity, possibly a highly degraded crater whose ejecta reaccreted onto the moon and created the smooth surface (Thomas et al., 1996; Thomas, 1998).

Although Phobos is low in density compared to mafic mineral assemblages and the bulk terrestrial planets, it is significantly denser than Deimos. Pätzold et al. (2013) used mass determined from Mars Express’ flyby of Phobos on March 3, 2010, at 77 km distance, and the global volume from Willner et al. (2010), to derive a bulk density of $1862 \pm 20 \, \text{kg m}^{-3}$. They were unable to solve for the second-degree gravity field to constrain internal heterogeneity. The best current mass for Deimos (Jacobson, 2010), divided by the best Viking-era volume estimate (Thomas, 1993), yields a bulk density of $1490 \pm 190 \, \text{kg m}^{-3}$, $2\sigma$ below Phobos’ density. In section 3.4 potential implications for the moons’ density difference are discussed. Both densities are lower than those of nearly all analog materials, and substantial internal porosity is probably present at least within Deimos (Rosenblatt, 2011; Murchie et al., 2013).

### 2.3. Gravitational Environment

Small asteroids and small bodies commonly have non-intuitive vectors for surface acceleration. These result from their irregular shapes; rapid spin, which enhances centripetal force; low mass, which decreases gravity; and especially in the case of Phobos, tidal effects. Determining what is “upslope” or “downslope” is best accomplished by calculating potential energies over the surface (Vanicek and Kakiwsky, 1986; Thomas, 1993). One measure of relative height, dynamical height, is obtained by dividing potential energy by acceleration. In the case of Phobos, as it spirals toward Mars, the longer ends (pointing toward and away from Mars, $0^\circ$ and $180^\circ$ longitude) become conspicuous gravitational lows due to tides and rotational accelerations (Fig. 3) (Thomas, 1993; Shi et al., 2012). Escape velocities near those locations are also highly directionally dependent, less than $3 \, \text{m s}^{-1}$ toward the east, but over $9 \, \text{m s}^{-1}$ to the west (Thomas, 1993). Deimos, orbiting much farther from Mars, experiences less effect from tides and rotational accelerations.

### 2.4. Regolith Properties

Having formed on small objects without atmospheres or internal activity, the regoliths of the two moons are expected to be the product of impacts and redistribution of their ejecta over geologic time. Nearly all ejecta from craters on Phobos and Deimos either reimpact the satellites directly [that fraction being a function of strength, which affects ejecta velocity (Asphaug and Melosh, 1993)], or escape to Mars orbit and reaccumulate over time (Soter, 1971; Ramsley and Head, 2014). An important question has been the abundance of Mars ejecta in Phobos’ regolith (Chappaz et al., 2013; Ramsley and Head, 2013a, 2014). Chappaz et al. (2013) and Ramsley and Head (2013a) calculated that Mars ejecta may constitute a few parts per million to a few hundred parts per million of Phobos’ regolith, most having accreted later in Phobos’ evolution as it has spiraled to lower altitudes. The presence of at least a few centimeters of low thermal inertia material, indicated by surface temperatures
determined from Viking Orbiter thermal infrared measurements (Lunine et al., 1982), is consistent with fine-grained, unconsolidated regolith.

Crater morphology reveals evidence for layering in Phobos’ regolith. Layering is directly exposed in one crater wall (Thomas, 1979). Crater floor morphologies [e.g., flat floors, concentric craters (Quaide and Oberbeck, 1968)] suggest depths to discontinuities of a few to several tens of meters in some areas of Phobos (Thomas et al., 2000). Basilevsky et al. (2014) used similar criteria to map crater morphology with Mars Express images, and found that the depth to discontinuities varies greatly over short distances. These and MARSIS radar sounder data (Heggy et al., 2014) suggest that Phobos’ regolith varies in modest ways horizontally and that it gradationally merges at greater depths with a more solid, but likely still porous object.

Large ejecta blocks are found on both moons. The largest on Phobos may be an ~85-m boulder near the crater Stickney; Deimos has a poorly resolved block ~150 m across. Basilevsky et al. (2014) noted that the surface density of large blocks (from a variety of source data) approximates that on some lunar terrains, and that mapping of both obvious blocks and mounds interpreted as blocks reveals an association with specific craters. Lee et al. (1986) found that large blocks are more widespread on Deimos than on Phobos, and typify the surface seen in the highest-resolution images (Fig. 3). The sizes of the large blocks are within ranges expected for the possible source craters, and likely were formed by similar excavation and fragmentation phenomena as lunar ejecta blocks. The abundance of smaller blocks on Phobos is known only for the region near Stickney imaged at high resolution, and it is comparable to the number density and size-frequency distributions of blocks on Eros and Itokawa (Ernst et al., 2015).

Phobos’ regolith exhibits color as well as albedo variations, whereas Deimos’ regolith has albedo variations but little color variation (Murchie et al., 1991; Thomas et al., 2011); both bodies are a factor of 2 lower in reflectance than even the most space-weathered lunar mare materials (sections 3.1, 3.2). Phobos’ greatest color variation is associated with the crater Stickney (Fig. 3), a “bluer” region that includes most of the crater, the ejecta to the east, and less distinct lobes to the northwest and southwest (Patsyn et al., 2012; Pieters et al., 2014). These locations correspond to predicted locations of the thickest ejecta from Stickney (Thomas, 1998). The rest of the surface is a “redder unit,” and close to the color exhibited by Deimos. Phobos has some brighter crater rims, and accumulations of relatively dark material on the floors of some craters. Deimos’ albedo features are mostly streamers that taper downslope from crater rims and blocks on the regional gravitational highs at ridge crests (Fig. 3) (Thomas et al., 1996). These albedo features exhibit little color contrast from the darker surroundings, but are up to 30% brighter than the rest of Deimos’ surface. It is likely that some of the elevated reflectances are related to lesser space weathering of more freshly exposed regolith, particularly on Phobos’ bright crater rims and in Deimos’ streamers. The detailed processes by which space weathering operates on Mars’ moons depends on the composition of the starting material, which is debated (section 3.2). However, it is unlikely that Phobos’ redder material results from space weathering of the bluer material: Space weathering is understood to weaken mineral absorptions (Hapke, 2001), and absorptions present at 0.65 and 2.8 μm in the redder material are weaker or absent in the bluer material (section 3.2).

Deimos exhibits clear evidence for widespread motion of material down regional slopes: brighter streamers trending down gravitational slopes, asymmetric ponded material within craters concentrated on their downslope sides, and more blocks and small craters on higher terrain (those at lower elevations having been buried). Deimos has perhaps the largest recognized crater of any small body relative to body size, the south polar concavity with a diameter of ~10 km (Thomas, 1989; Morrison et al., 2009). The estimated volume of this crater is 43 ± 20 km³ on an object with a surface area 520 km². In comparison, Stickney’s initial volume was 50 ± 15 km³ on an object with a surface area of 1600 km² (Thomas, 1998). Nearly all crater ejecta should have eventually reaccreted on Deimos. At present, an uneven distribution of craters larger than 350 m indicates uneven blanket or other degradation of craters. The simplest explanation, consistent with the topographic slopes, is concentration of reaccreted ejecta into lower areas by mass wasting, yielding thicknesses up to 200–400 m in gravitational lows (Fig. 14 of Thomas, 1998). Deimos’ globally uniform appearance suggests that the south polar cratering event does not date to Deimos’ earliest history: If the cratering event was that old, then downslope movement of regolith might have exposed substrate in topographically higher regions, which would then have accumulated a high density of small impact craters.

Whereas Deimos exhibits mass wasting on slopes as low as 2°, Phobos’ rougher surface has only localized examples of mass wasting. The most obvious are on the western interior slope of Stickney (Thomas et al., 2000; Basilevsky et al., 2014), which is at the angle of repose and has few superposed small craters and numerous downslope albedo streamers. On the floor of Stickney is a hummocky mound several hundred meters in thickness that is probably a debris slide formed substantially after Stickney’s formation, judging from its well-preserved hummocks and few superposed craters. As noted in section 2.3, the gravitational slopes on Phobos are changing as it approaches Mars; Stickney’s western slope has probably increased with time, encouraging further and future mass movement. Other mass movement on Phobos is observed in smaller craters (Basilevsky et al., 2014) and in some grooves.

Both martian moons have been observed by Earth-based radar. Busch et al. (2007), using Arecibo observations at 13-cm wavelength, measured radar albedos of 0.056 ± 0.014 for Phobos, consistent with a recalibrated value of 0.049 ± 0.01 for Phobos found by Ostro et al. (1989), and 0.021 ± 0.006 for Deimos. Both radar albedos are low: Phobos is near the low end of asteroid radar albedos and Deimos
has the lowest detected radar albedo in the solar system to date. Given the radar albedos, a near-surface density and porosity can be calculated if a composition is assumed; Busch et al. (2007) considered lunar soils and dehydrated low-grade carbonaceous chondrites, and for each used grain densities of 2700 kg m$^{-3}$. Phobos' near-surface bulk density was estimated to be 1600 ± 300 kg m$^{-3}$, indicating a near-surface total porosity of 40 ± 10%, while for Deimos the values are 1100 ± 300 kg m$^{-3}$ and 60 ± 10%.

2.5. Crater Densities and Surface Ages of the Moons

Crater densities on Phobos have been studied by Thomas and Veverka (1980), Thomas (1998), and Schmedemann et al. (2013, 2014). Studies of crater density on irregular objects must deal with geographic variations in image quality and lighting, effects of differing slopes, and relief of the substrate that can affect crater shape and size. Crater degradation on small objects may involve seismic shaking induced by impacts (Richardson et al., 2005). Schmedemann et al. (2014) found spatial variations in crater density on Phobos, and possibly discontinuities in the cratering record (kinks in the cumulative crater density curves), which they hypothesized to be created by seismic shaking or ejecta emplacement events. Absolute ages are model-dependent: If Phobos has experienced the expected crater flux at Mars’ orbit since its earliest history, then the oldest surface may approach 4.3 Ga and Stickney may be as old as 4.2 Ga. Model ages for grooves are younger than Stickney, consistent with the observation that grooves cross-cut Stickney and its ejecta, but their absolute age probably exceeds 3 Ga. The role of secondaries is unclear: Ramsley and Head (2014) interpreted it to be significant, whereas Schmedemann et al. (2014) did not. The antapex (around 270°W longitude) lacks significant grooves and has been suggested to have less regolith than other areas (Thomas, 1998), perhaps relating both to deposition of ejecta near Stickney, and to disturbance near Stickney’s antipode. A global crater catalog of Phobos has been compiled, and is likely close to complete for craters >50 m in diameter (Bandeira et al., 2014).

Deimos’ distinct morphology indicates a different history of crater modification from that on Phobos. Despite smoothing of the surface by deposition or reaccumulation of a large amount of ejecta (section 2.4), there remains a high density of large craters (Thomas, 1998) not very different from that on Phobos. However, although rims of larger craters are visible for counting, crater interiors have been largely filled by debris (Thomas, 1998). The uneven geographic distribution of smaller craters on Deimos, with fewer small craters near the center of the topographic facets, indicates greater burial by debris in gravitational lows (Fig. 16 of Thomas, 1998).

2.6. Grooves on Phobos

Viking Orbiter views of Phobos revealed linear, often parallel depressions 100–200 m wide, extending up to several kilometers (Veverka and Burns, 1980). Several spacecraft missions later, these features remain without consensus explanations. Similar but not identical features have been found on small icy satellites (Morrison et al., 2009), and on small asteroids including (951) Gaspra (Veverka et al., 1994), (243) Ida (Sullivan et al., 1996), (433) Eros (Buczkowski et al., 2008), and (21) Lutetia (Thomas et al., 2012). A large fraction of Phobos’ grooves fall into at least four parallel sets, whose normals are close to the plane defined by the long and short axes of Phobos (Thomas et al., 1979; Murray et al., 1994). Some grooves cut other grooves, and many cut crater rims, including Stickney’s. They traverse a wide range of angles with respect to local slope directions. Many have a pitted appearance, although some have straight rims with no apparent pitting. Many are “shoulder-to-shoulder.” Raised rims appear to bound some grooves, although the shoulder-to-shoulder groove spacing in some places and subtle reflectance variations make rim topography difficult to characterize. Interior slopes are typically well below angles of repose. A few wide, shallow grooves are the clearest examples of linear arrays of discrete pits (D grooves of Thomas et al., 1979).

Hypotheses for the origin of the grooves include secondary cratering from Phobos impacts (Veverka and Duxbury, 1977; Head and Cintala, 1979), secondary cratering from Mars impacts (Murray et al., 1994, 2014), cratering from debris in Mars orbit (Schultz and Crawford, 1989; Kikuchi and Miyamoto, 2014), tracks formed by rolling ejecta blocks (Wilson and Head, 2005; Duxbury, 2011; Hamelin, 2011; Ramsley and Head, 2013b), and fractures of the body of Phobos into which regolith has drained or from which regolith has been ejected by degassing of water from the interior (Thomas et al., 1979; Weidenschilling, 1979; Horstman and Melosh, 1989). The grooves show a variety of morphologies, which could be evidence that their formation involved more than a single mechanism. The range of groove morphologies observed in different materials and dynamical environments of asteroids and icy satellites (Buczkowski et al., 2008; Murray et al., 2009; Thomas et al., 2012) suggests that multiple formation mechanisms act throughout the solar system. One common feature in all models of groove formation, however, is a requirement for loose regolith, to tens of meters in depth, where grooves occur.

None of the hypotheses for groove formation is overwhelmingly supported for all of Phobos’ grooves by existing observations. Secondary cratering from Mars theoretically could explain some grooves’ alignments (Murray et al., 1994), but the close spacing of the grooves, the even spacing of similarly sized pits along the grooves, and the large area of coverage by grooves cannot obviously be explained by secondary cratering from impacts many thousands of kilometers away (Ramsley and Head, 2013b). A few distinct, wide grooves do have morphologies consistent with secondary crater chains, and have orientations that match predicted ejecta patterns from Stickney (Thomas, 1998), but most grooves do not. Impacts from debris in orbit about Mars might be consistent with some patterns (Schultz and Crawford, 1989; Kikuchi and Miyamoto, 2014), but this hypothesis is not well developed theoretically. Rolling
ejecta blocks can form somewhat similar aligned markings, but there is no explanation for the consistent parallel patterns of grooves instead of meandering paths deflected by topography that are typical of lunar boulder tracks, or for the paths traversing all slope angles [especially apparent east of Stickney; see Fig. 10 of Thomas (1998)]. Some grooves, especially those crossing the northern polar region, have morphologies (straight walls, nearly constant widths) suggestive of graben, as do some grooves on saturnian satellites with similar body-related patterns (Morrison et al., 2009). Other grooves consist of regularly spaced chains of similarly-sized pits consistent with drainage of loose regolith into underlying fractures (Horstman and Melosh, 1989). Tidal stretching has been a favored source of stress to generate fractures underlying grooves (Soter and Harris, 1977; Weidenschilling, 1979; Morrison et al., 2009). Mechanical modeling based on the real shape of Phobos and on a layered or gradational regolith is needed to test a possible origin by fracturing.

2.7. Surface and Orbital Environment

The surface environments of Phobos and Deimos are strongly affected by their location in Mars orbit. At the surface, at noon near the subsolar point, surface temperature may exceed 350 K at perihelion, but only reach 270 K when Mars is at aphelion (Giuranna et al., 2011). Modeling predicts that nighttime surface temperature depends on latitude and season, typically cooling to 100 K before dawn, although temperatures during polar night may reach 60 K. The diurnal thermal wave penetrates to less than 1 cm, and at depth the average temperature is near 230 K (Kuhrt and Giese, 1989). Unlike the Moon and Mercury, Phobos and Deimos do not have polar craters that are permanently shadowed and ≤100 K in temperature over geologic time. On geologic timescales, Mars’ obliquity is chaotic, ranging between 0° and 60° (Laskar and Robutel, 1993). For the moons’ low orbital inclinations to Mars’ equator, orbit inclination reckoned from Mars’ equator of date is subject only to small variations — in other words, as Mars’ obliquity varies, the moons’ orbits track the equator (Gurfil et al., 2007). Thus, presently cold polar regions were subjected to prolonged heating by insolation in the past (Laskar et al., 2004). This history also affects the moons’ ability to have retained internal ice that has been hypothesized to have been present at the moons’ formation (e.g., Rosenblatt, 2011). At all latitudes, any initial pore-filling ice will have ablated to a depth of between hundreds of meters to over 3 km, depending on typical pore size (Fanale and Salvail, 1989).

The regolith environment may be even more exotic, with the moons predicted by theory to reside in a “dust belt” (Soter, 1971). No direct measurement or even confirmation of such a belt exists, so the details of dust belt density and morphology are based on models. Observations by the Hubble Space Telescope place upper limits on the optical depths (Showalter et al., 2006). Modeling predicts that much of the ejecta from small impacts on the moons will exceed the escape velocity of either moon and remain in Mars orbit. Larger particles will follow Keplerian trajectories and eventually reimpact the parent moon, whereas the orbits of smaller particles will be perturbed by solar radiation pressure, and become asymmetric about Mars. Small particles originating from Phobos are perturbed toward the Sun, and those originating from Deimos are perturbed away from the Sun. In addition, grains from Deimos will oscillate about size-dependent Laplace planes, creating a vertically asymmetric ring. For the Deimos ring, interaction between the two asymmetries gives rise to a time-dependent structure that varies with martian season. The smaller particles eventually either impact Mars or are lost from the Mars system (Horanyi et al., 1990, 1991; Kholshevnikov et al., 1993; Juhasz and Horanyi, 1995; Sasaki, 1996; Hamilton, 1996; Krivov and Hamilton, 1997; Makuch et al., 2005; Krivov et al., 2006; Zakharov et al., 2014).

3. COMPOSITION, ORIGIN, AND ORBITAL EVOLUTION

The compositions of Phobos and Deimos are interesting in themselves, but more importantly, they help to constrain the moons’ origins. At the present time, remote measurements can provide two major constraints on the moons’ composition, their overall reflectance and spectral continuum, and the presence of absorption or emission features diagnostic of particular minerals.

3.1. Constraints from Reflectance

Comparisons of the visible to near-infrared reflectances of Phobos and Deimos with analog materials have been hampered over the years by comparison of dissimilar quantities, e.g., geometric albedo and bidirectional reflectance measured in the laboratory, or single-wavelength measures of reflectance that do not include an assessment of the spectral continuum. Mars Express/OMEGA and MRO/CRISM observations have overcome these challenges because they cover hundreds of wavelengths, and the OMEGA data cover a range of geometries that allows derivation of an accurate photometric correction to the geometry at which analog materials are commonly observed in the laboratory, i.e., incidence angle \( i = 30° \), emergence angle \( e = 0° \), and phase angle \( \alpha = 30° \) (Fraeman et al., 2012). Figures 4a,b compare corrected spectral reflectance of Phobos to examples of the two broad classes of proposed composition, carbon-rich materials exemplified by low-grade carbonaceous chondrites, and space-weathered mafic assemblages exemplified by mature lunar mare soils. Reflectance and spectral slope are closely comparable to low-grade, carbon-rich CI and CM carbonaceous chondrites, and are particularly well matched by samples of the CM chondrite Murchison that have been heated by laser pulses to simulate the effects of space weathering (Hiroi et al., 2003). Pajola et al. (2013) were able to model the spectrum of the redder unit by adding a component of Mg-rich glass to the Tagish Lake
ungrouped, phyllosilicate-rich carbonaceous chondrite. In contrast with the close match in reflectance and spectral slope with primitive carbonaceous materials, Phobos and Deimos are a factor of 2 darker than even the most space-weathered lunar mare soils, making space-weathered Mars crustal material unsuitable as a spectral analog despite the similarity in spectral continuum slope. Phobos and Deimos are also much darker, and much redder, than shock-darkened ordinary chondrites, which have been suggested as compositional analogs to several low-reflectance planetary surfaces (Britt and Pieters, 1988; Gillis-Davis et al., 2013).

### 3.2. Constraints from Spectral Features

Constraining the compositions of Phobos and Deimos from weak mineral spectral signatures requires accurate calibration of remote spectral measurements, confirmation of the mineral signatures in multiple datasets, and determining the diagnostic character of the signatures. In ISM spectra taken from Phobos 2, upward curvature in the 1-µm region in a ratio spectrum of bluer to redder material resembles an absorption due to olivine or possibly pyroxene. Such a feature would be consistent with less-red material having experienced less lunar-like space weathering and having a stronger mineral signature (Murchie and Erard, 1996; Gendrin et al., 2005). Disk-integrated observations of the moons acquired from Mars’ surface by the Imager for Mars Pathfinder showed no evidence for the 1-µm feature, and indicated instead a broad, shallow feature near 0.7 µm (Murchie et al., 1999) comparable to that seen in some low-albedo asteroid spectra (Vilas et al., 1994). Subsequent OMEGA and CRISM data do not reproduce detection of a 1-µm absorption due to olivine or pyroxene (Fraeman et al., 2012, 2014) or detect a feature due to H2O at 3 µm (Fraeman et al., 2012).

However, CRISM data do reveal spectral features in both units (Fraeman et al., 2014). Deimos and the redder unit on Phobos both exhibit a broad, shallow absorption centered near 0.65 µm — the feature detected earlier in Mars Pathfinder data — but the feature is absent from Phobos’ bluer unit. The strength of this feature is highly correlated with visible color observed by HiRISE and CRISM. The 0.65-µm

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**Fig. 4.** CRISM spectral reflectance measurements of Phobos and Deimos. Uncertainty in the data is typically ±1% near 0.7 µm and ±3% near 3 µm (Fraeman et al., 2014). (a) Reflectance of bluer unit corrected to $i = 30^\circ$, $e = 0^\circ$, $\alpha = 30^\circ$ compared with meteorite analogs measured at the same geometry. Stickney ejecta and heated Murchison overplot each other. Adapted from Fraeman et al. (2012). (b) Redder and bluer units compared with mature lunar mare soils. Is/FeO is magnetic susceptibility divided by Fe content expressed as FeO, and is proportional to the extent of space weathering. Adapted from Fraeman et al. (2012). (c) CRISM spectra of redder and bluer units covering the 0.65-µm feature, compared with telescopic spectra that reproduce the absorption. Adapted from Fraeman et al. (2014). (d) Spectra corrected for thermal emission extending to 3.1 µm, showing the 2.8-µm OH absorption. Adapted from Fraeman et al. (2014).
absorption is reproduced in telescopic spectra dominated by the redder unit, but absent in telescopic spectra dominated by the bluer unit (Fig. 4c). Both units also exhibit a sharp absorption near 2.8 μm (Fig. 4d) attributable to OH (Clark et al., 1990); the feature is strongest on Deimos and weakest in Phobos’ bluer unit. Spectrally, Phobos and Deimos belong to the “Pallas group” described in the chapter by Rivkin et al. in this volume. With more complete wavelength coverage, the false detection of olivine from ISM spectra is now recognizable as the artifact of ratioing a bland bluer unit spectrum to a redder unit spectrum with a 0.65-μm absorption having a longer-wavelength shoulder near 1 μm.

One interpretation of these absorptions, consistent with the reflectance and spectral slope characteristics described in section 3.1, is a composition rich in phyllosilicate that has been desiccated of molecular H₂O. In particular, phyllosilicate-rich CM carbonaceous chondrites exhibit an OH feature like that observed at 2.8 μm as well as an absorption due to Fe in phyllosilicate near 0.7 μm (Cloutis et al., 2011; Takir et al., 2013; see also the chapter by Rivkin et al. in this volume). Phobos’ and Deimos’ Fe feature is centered closer to 0.65 μm, possibly suggesting a different desiccated phyllosilicate, such as Fe-bearing nontronite (Fraeman et al., 2014). That interpretation would be consistent with an origin of the moons that includes primitive material. However, alternate interpretations involving anhydrous minerals consistent with other proposed origins cannot be entirely excluded. Curvature in the spectrum near 0.65 μm might result from Rayleigh scattering by a mixture of microphase and nanophase Fe in the right proportion that is formed by space-weathering processes (Clark et al., 2012). The OH feature near 2.8 μm is similar in its reflectance minimum to an OH feature observed on the Moon (Clark, 2009; Pieters et al., 2009; Sunshine et al., 2009) and attributed to anhydrous reaction of implanted solar wind protons with oxygen in silicate minerals (McCord et al., 2011), although the lunar feature is considerably broader than that observed on Phobos, Pallas, or CM meteorites.

Thermal infrared spectra complement the mineral detection capabilities of visible-to-near-infrared spectra. However, their analysis is complicated by the need to ratio to a blackbody spectrum, and the fact that Phobos’ and Deimos’ thermal emissions are not well fit by an ideal blackbody (Lynch et al., 2007). Giuranna et al. (2011) calculated emissivity spectra from TES and PFS data whose footprints cover large segments of Phobos, assuming an areal mixture of blackbody spectra representing three different surface temperatures (shaded surfaces, low and high thermal inertia solar-illuminated surfaces). Selected regions exhibit apparent emissivity features consistent with phyllosilicate, although features in other spectra resemble those due to feldspathoid minerals. Giuranna et al. favored the feldspathoid interpretation, and a composition of highly evolved Mars crustal material. Difficulties in this interpretation include the lack of feldspathoid-rich meteorite analogs, the dominance of basaltic compositions in Mars’ crust (McSween et al., 2009), the lack of detection of feldspathoids on Mars despite detection of over 30 other minerals (Viviano-Beck et al., 2014), and the possibility of artifacts from assuming a mixture of surfaces with discrete temperatures. More recently, using TES data, Glotch et al. (2015) modeled thermal emissivity from TES data covering the redder unit using a surface with many different temperatures. They find evidence for bound molecular water and carbonate, which if correct would be diagnostic of primitive carbonaceous compositions, and for silicate consistent with the desiccated phyllosilicate inferred from visible to near-infrared reflectance.

Regardless of precise compositions, the contrast in spectral properties between the bluer and redder units provides information on three-dimensional structure within Phobos. Inside Stickney, either redder or bluer material is exposed in different locations (Thomas et al., 2011). The heterogeneity inside Stickney led Basilevsky et al. (2014) to suggest that at depth Phobos consists of a rubble of discrete bluer and redder blocks as proposed by Murchie et al. (1991). Alternatively, the interior may be relatively coherent as suggested by average density if a CM-chondrite-like composition is assumed (section 3.4), with the bluer material that is relatively “pristine” chemically having zones and pockets of hydration into redder material richer in phyllosilicates, which exhibit 0.65-μm and 2.8-μm absorptions.

3.3. Comparison with D-Type and Related Asteroids

The spectra of martian satellites are consistent with low-reflectance, outer-belt D-class asteroids, which also dominate the Trojan asteroid region. This similarity was first noted for Deimos by Grundy and Fink (1992), and for Phobos by Murchie and Erard (1996), who found that most of Phobos’ surface has a spectrum similar to that of Deimos, although the area near Stickney has a flatter spectral slope. Rivkin et al. (2002) combined 1.65–3.5-μm spectra from the Infrared Telescope Facility (IRTF) with shorter-wavelength Phobos 2 data, and interpreted D-class asteroids to best match Deimos and Phobos’ redder unit, whereas Phobos’ bluer unit is more similar to P- or T-class asteroids.

The weak absorptions observed on Phobos and Deimos are also consistent with those observed on low-reflectance asteroids. Absorptions near 0.7 μm on low-reflectance asteroids have been attributed to Fe-bearing phyllosilicates, which are usually accompanied by absorptions due to OH or H₂O near 3 μm (Vilas and Gaffey, 1989; Vilas et al., 1993; Vilas, 1994; Rivkin et al., 2002). Mineralogical interpretations for the asteroids most similar to Phobos and Deimos having absorptions near 0.65 μm without an accompanying feature near 0.9 μm remain uncertain (Jarvis et al., 1993; Vilas et al., 1994). Many low-reflectance asteroids also display an absorption centered near 2.8 μm, and these features can be grouped qualitatively into three classes (see the chapter by Rivkin et al. in this volume): (1) (1) Ceres-type, with symmetric absorption minima near 3.0 and 3.3 μm, possibly attributable to brucite (Milliken and Rivkin, 2009); (2) (24) Themis-type, with an absorption minimum near 3.1 μm attributable to water frost (Rivkin and Emery, 2010; Campins et al., 2010); and
(3) (2) Pallas-type, having a checkmark-shaped absorption with an minimum near 2.8 µm attributable to OH-bearing phases (Rivkin et al., 2011). Phobos’ and Deimos’ features group with those of the (2) Pallas class.

3.4. Implications for Internal Structure

Given a compositional determination for either Phobos or Deimos, and a measurement of the moon’s average density, the approximate internal structure of the moon can be inferred. As the extent of fracturing of an asteroidal body increases and its mechanical coherence decreases, internal structure transitions through four classes described by Wilkison et al. (2002) and Brit et et al. (2002): coherent, with minimal macroporosity; fractured but mechanically coherent, with up to 20% macroporosity; heavily fractured, with 20–30% macroporosity; and loosely consolidated or rubble pile, with ≥30% macroporosity. For small bodies whose internal pressures are smaller than the mechanical strength of the constituent materials, additional porosity (microporosity) occurs as pores or microfractures between individual mineral grains within mechanically coherent chunks. The macroporosity is needed to infer interior structure. [Definitions of density and porosity of meteorites and asteroids used here follow those of Brit and Consolmagno (2003) and Consolmagno et al. (2008).] Grain density is defined as the density of the solid phase of a meteoritic or asteroidal material, excluding voids. Bulk density is the density of a porous meteoritic material, and differs from grain density due to the presence of pores and microfractures between grains. Macroporosity is the volume percent of an asteroidal body occupied by fractures between mechanically coherent chunks, not including pores between grains within those chunks (microporosity). Total porosity of an asteroidal body is the sum of macroporosity and microporosity.

Inferred internal structures inferred for Phobos and Deimos depend strongly on the composition assumed. CM meteorites, arguably the closest spectral analog to the moons, have a bulk density (including pores and microfractures) of 2250 ± 80 kg m⁻³, and a grain density (for the solid phase only) of 2900 ± 80 kg m⁻³ (Consolmagno et al., 2008). Using this CM bulk density, Murchie et al. (2013) calculated a macroporosity for Phobos of 17 ± 4%, placing it in the fractured but coherent category, and a macroporosity for Deimos of 34 ± 11%, consistent with a rubble pile. A more porous structure for Deimos would be consistent with its south polar concavity having originated from an impact that partially shattered the moon, followed by reaccretion of the fragments to explain the smooth surface (Thomas, 1996). The derived macroporosity for Deimos is comparable with typical values for fragmental lunar mare regolith at tens of centimeters to meter depth, 35–38% (Mitchell et al., 1972), where overburden pressure is comparable to that in the martian moon interiors. Andert et al. (2010) used a slightly lower grain density to estimate a total porosity for Phobos of 30 ± 5%. Andert et al. equated this result with a loosely consolidated interior, but that interpretation did not allow for partitioning of total porosity into microporosity and macroporosity, only the latter of which is relevant for internal structure.

Rosenblatt (2011) considered what macroporosities would correspond with differing compositional interpretations of existing spectral measurements. For example, if the moons are dominated by Mars crustal material (Cradlock, 2011; Giuranna et al., 2011) or shock-darkened ordinary chondrite (Britt and Pieters, 1988), macroporosity for Phobos could be in the range of 25–47% and for Deimos in the range 41–58%. Rosenblatt speculated that the higher porosities would be unrealistic and require portions of the void spaces to be filled by water ice. However, an explanation for the geochemically unexpected combination of anhydrous silicates with a large fraction of ice — two components expected to have condensed in different parts of the solar system — was not proposed.

Interpretation of geologic surface features on the martian moons, and in particular the grooves on Phobos, also has the potential to provide insights into internal structure. The formation mechanisms for Phobos’ grooves are actively debated, as discussed in section 2.6. Models that invoke extensive fracturing, if correct, would indicate mechanical coherence consistent with that implied by a CM-chondrite-like composition. In contrast, continuous fracturing over kilometers would be inconsistent with the rubble-pile structure required by mafic or ordinary chondritic compositions.

3.5. Orbital Evolution and Formation Mechanisms

Any hypothesis for the origin of Phobos and Deimos must be consistent with the dynamics of the moons’ orbital evolution. Reconciling empirical observations of the moons’ properties with predictions of orbital dynamical models has proven challenging. A primitive carbonaceous composition, if that interpretation is correct, could have resulted from formation in the solar nebula at heliocentric distances beyond Mars’ orbit (Pollack et al., 1978), migration toward the inner solar system as proposed by some dynamical models [namely the Nice model (Tsiganis et al., 2005; Gomes et al., 2005)], and capture into Mars orbit. Capture requires special circumstances such as drag from an extended protoatmosphere (Hunten, 1979; Pollack et al., 1979; Sasaki, 1990).

Circularizing an initially eccentric orbit after capture, and reducing orbital inclination to the present near-equatorial orbit, can be achieved by tidal processes in the case of Phobos over the last 4.6 G.y. However, this requires a high tidal dissipation rate that is difficult to reconcile with a largely coherent Phobos. Instead, the interior would need to be highly fragmented, or have a substantial fraction of ice to increase the tidal dissipation rate (Lambeck, 1979; Mignard, 1981; Rosenblatt, 2011; McCarthy and Castillo-Rogez, 2011). In the case of Deimos, tidally induced orbital changes are probably too slow to account for evolution of an eccentric, inclined orbit over the last 4.6 G.y. (Lambeck, 1979; Szeto, 1983), unless there was enhanced tidal dissipation due to internal ice (Rosenblatt, 2011). Alternatively, drag in the
protoatmosphere, particularly efficient in circularizing an orbit, may have played a role. This scenario requires a very early capture, a suitable density of the protoatmosphere to prevent the captured asteroid from impacting Mars (Sasaki, 1990), and an extent of the protoatmosphere beyond the current orbit of Deimos.

The moons’ present orbits are thought to have evolved considerably once in the equatorial plane. Because Phobos orbits inside a synchronous orbit, it exhibits acceleration along its orbit resulting from the solid body tides raised by Phobos in Mars (e.g., Burns, 1992), by about $1.27 \times 10^{-3}$ deg yr$^{-2}$ (Lainey et al., 2007; Jacobson, 2010). Phobos’ orbit thus decays by about 20 cm yr$^{-1}$. Former moons orbiting well inside Phobos already will have impacted Mars. The interpretation of elongated martian impact craters being formed by obliquely impacting former moons is debated (Schultz and Luz-Garrihan, 1982; Botte et al., 2000; Chappelow and Herrick, 2008). The same mechanism is expected to cause Deimos’ orbit to recede away from Mars, because Deimos orbits beyond a synchronous orbit. When Phobos’ and Deimos’ orbits are integrated backward in time, inevitably the moons must once have been in close proximity, raising the possibility that they both formed near a synchronous orbit (Rosenblatt and Charnoz, 2012), or even are fragments of a single captured object (Singer, 2007; Rosenblatt, 2011).

Formation of Phobos and Deimos in orbit around Mars avoids the challenges of explaining the orbital evolution of captured objects. Phobos and Deimos could have coaccreted with Mars (Safonov et al., 1986), which would account for the current near-equatorial and near-circular orbits of both moons. However, this hypothesis implies that Phobos and Deimos would have ordinary-chondrite-like or bulk-Mars-like compositions, which is hard to reconcile with their low densities and spectral properties. One hypothesis that may be reconcilable with many observed properties of the moons is that they formed from a hot, circumplanetary accretion disk formed by the impact of a large carbonaceous body on Mars (Craddock, 2011; Rosenblatt and Charnoz, 2012; Canup and Salmon, 2014). Crater scaling relations identify the Daedalia, Chryse, and newly recognized Borealis (Andrews-Hanna et al., 2008) basins on Mars as the most plausible records of that impact (Craddock, 2011). The 4.3-Ga estimated crater age of the Borealis basin (Frey, 2010) agrees with the observed crater density on Phobos (Schemedemann et al., 2014). Material from the impactor could have dominated the accretion disk. Even if a carbonaceous impactor caused the moons’ low reflectances, heating of the material should have driven off molecular water, explaining the lack of a strong 3-μm absorption at present (Craddock, 2011). Rosenblatt and Charnoz (2012) examined several different scenarios for satellite formation from such an accretion disk, in a strong-tide regime close to the Roche limit and in a weak-tide regime further out. They concluded that formation in a strong-tide regime would have formed bodies whose dynamic lifetime against reimpact on Mars is much less than the moons’ age, so that such a mechanism cannot have formed the moons. In contrast, formation in a weak-tide regime between the Roche limit and synchronous orbit may have been capable of forming the satellites. They suggested that Phobos and Deimos may be the last two remnants of such bodies formed near the synchronous distance to Mars, other now extinct moons having since impacted Mars.

4. OUTSTANDING QUESTIONS FOR FUTURE STUDY

After Mars Express, four key uncertainties remain about Phobos and Deimos for which resolution is required to answer high-level science questions. First, what are the moons’ compositions, and how does that constrain the moons’ origins? Some remote measurements are highly suggestive of a CM-carbonaceous-chondrite-like composition, but alternative interpretations of the moons’ low reflectance and weak spectral features cannot be ruled out based on existing data. Sample return answers this question definitively using major and minor elements and isotopes to link the moons to particular reservoirs. Short of sample return, future remote or in situ spacecraft measurements could also provide key compositional information to address the fundamental question of the moons’ origins. As shown in Table 2, a diagnostic set of measurements would be elemental abundances, particularly measurements that would constrain abundances of H, C, S, and major and minor elements exhibiting a range of volatilities (e.g., Zn, Mn, and Al). Another diagnostic measurement set would be of mineralogy with a focus on phases whose abundances differ greatly between proposed compositions, particularly olivine, pyroxene, phyllosilicate, and carbonate. The ambiguity of detection of these phases in existing measurements suggests that new measurements should include fresh, non-space-weathered surfaces, or should resolve single mineral grains to enhance spectral signatures of minor phases. The presence of molecular water, hydroxylated minerals, or subsurface ice is a critical test of models of the moons’ formation: A high content of water or hydroxyl in minerals (or water-equivalent H) is consistent with capture of a primitive body, but is probably inconsistent with formation in situ. For in situ formation, either the source material was low-water (ordinary chondrite, martian crust) or it accreted from a circumplanetary disk from which water was baked out.

Second, what is the relationship of Phobos to Deimos? In part this depends on their origins — if formed in Mars orbit, they are related, but if captured, they could be either unrelated or fragments of a single body that was disrupted during capture (Singer; 2007; Rosenblatt, 2011). In addition, the moons’ closer proximity earlier in their history and regolith exchange between the moons via the hypothesized dust belts raises the possibility that Phobos’ redder unit could be derived from Deimos. Testing these hypotheses requires modeling as well as compositional measurements of both moons.

Third, what is the moons’ deep structure and could there be deep water ice? The two most important new groups of measurements would be composition of both moons to estimate internal macroporosity, and volume of Deimos to
reduce the uncertainty in its average density. Higher-degree gravity coefficients for both moons would also be useful. Given the likelihood that any internal ice has ablated to a depth of 1 km or greater, near-surface detection of it by shallow radar sounding may not be feasible. Alternatively, heterogeneity in the filling of deep pores by water ice may be detectable by measuring the amplitude of Phobos’ forced libration and the second-order coefficients of the nonspherical part of the gravity field (Rosenblatt et al., 2011). Measurement of the forced libration can also be used to test for enhanced macroporosity beneath Stickney due to impact fracturing (Rambaux et al., 2012; LeMaistre et al., 2013).

Fourth, what is the origin of the grooves on Phobos? The global distribution, orientation, and stratigraphic relations of grooves is now relatively well known, but implications of these observables for models of groove origin remain debatable (Ramsley and Head, 2013b; Murray and Heggie, 2014). Given the widespread occurrence of somewhat similar, but varied features on asteroids and on small icy bodies, understanding the way or ways in which grooves form is important for interpreting the evolution of small bodies across the solar system. New tests for models of groove origin may be obtained using high-resolution, meter-scale measurements of groove topography and color. Various models make different predictions about topography of groove rims and floors, and about stratigraphic relations of groove floor and rim materials. Seismic measurements across grooves could determine whether they are surface expressions of fractures at depth.

Many key outstanding scientific questions about the martian moons may be solved by future robotic missions, even before ultimate human exploration of these destinations.

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