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The origin of the Martian moons revisited

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Abstract The origin of the Martian moons, Phobos and Deimos, is still an open issue: either they are asteroids captured by Mars or they formed in situ from a circum-Mars debris disk. The capture scenario mainly relies on the remote-sensing observations of their surfaces, which suggest that the moon material is similar to outer-belt asteroid material. This scenario, however, requires high tidal dissipation rates inside the moons to account for their current orbits around Mars. Although the in situ formation scenarios have not been studied in great details, no observational constraints argue against them. Little attention has been paid to the internal structure of the moons, yet it is pertinent for explaining their origin. The low density of the moons indicates that their interior contains significant amounts of porous material and/or water ice. The porous content is estimated to be in the range of 30–60% of the volume for both moons. This high porosity enhances the tidal dissipation rate but not sufficiently to meet the requirement of the capture scenario. On the other hand, a large porosity is a natural consequence of re-accretion of debris at Mars' orbit, thus providing support to the in situ formation scenarios. The low density also allows for abundant water ice inside the moons, which might significantly increase the tidal dissipation rate in their interiors, possibly to a sufficient level for the capture scenario. Precise measurements of the rotation and gravity field of the moons are needed to tightly constrain their internal structure in order to help answering the question of the origin.

Keywords Planets and satellites: formation · Planets and satellites: interiors · Planets and satellites: individual (Mars, Phobos, Deimos)

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1 Introduction

Unlike for the Earth's moon, the origin of the two small moons of Mars is still an open issue in spite of numerous spacecraft missions sent to the Martian system. It has been proposed that both moons were formed away from Mars' orbit and then were captured by Mars' gravitational attraction (e.g. Burns 1992) or that both moons were formed in situ from a circum-Martian disk of debris (e.g. Peale 2007). The capture scenario is mainly based on the striking similarities between the physical characteristics of the surface of Phobos and Deimos and those of numerous small-sized objects of the main or outer part of the asteroid belt (e.g. Pollack 1977; Thomas et al. 1992). In particular, the matching of the reflectance spectra of these surfaces with those of low-albedo asteroids hints at a carbonaceous chondrite composition for both moons (e.g. Pang et al. 1978; Pollack et al. 1978; Murchie et al. 1991; Rivkin et al. 2002), suggesting that Phobos and Deimos formed in the solar nebula at heliocentric distances beyond Mars' orbit (Pollack et al. 1978; Burns 1992). However, no satisfactory meteorite spectral analog to Phobos and Deimos has been found so far (e.g. Murchie et al. 1991; Murchie and Erard 1996; Vernazza et al. 2010). In addition, the capture scenario has major difficulties to account for the current near-equatorial and near-circular orbit of the moons around Mars (see the review of Burns 1992).

On the other hand, these orbits are consistent with expected orbits of objects accreted around Mars (Safronov et al. 1986), thus motivating some authors to propose that Phobos and Deimos were formed in Mars' orbit instead of being outer solar system objects captured by Mars. Some of these scenarios are not inconsistent with a possible carbonaceous composition for Phobos and Deimos, since they may involve the entering of an object from the outer part of the solar system (so, possibly of carbonaceous composition) into orbit around Mars, from which Phobos and Deimos would have been formed after the destruction of this object either by tidal forces from Mars (Singer 2003, 2007) or by collision with Mars (Craddock 1994, 2011). Moreover, among the small moons of the giant planets, some are thought to be captured objects (for instance, Saturn's moon Phoebe (Johnson and Lunine 2005) or Jupiter's moon Almathea (Anderson et al. 2005)) while others are thought, in the case of Saturn, to be objects accreted in situ from the rings of the planet (Charnoz et al. 2010). In turn, it suggests that, in spite of their similarities with some small bodies of the solar system, small moons of planets have not necessarily been captured by these planets.

In all the studies about the origin of the Martian moons, little attention has been paid to their internal structure and the possible link with their origin. The structure and properties of the interior of these bodies provide valuable information about the physical processes prevailing at their origin. For example, the knowledge of the dissipative properties of the interior is key for the understanding of the past evolution of the orbit of Phobos and Deimos (e.g. Lambeck 1979; Mignard 1981), which is directly relevant to the challenges raised by the capture scenario (Burns 1992). The interior of the Martian moons was, unfortunately, poorly constrained by historical observations, performed by the former Viking and Phobos-2 missions (Dobrovolskis 1982; Avanesov et al. 1991; Murchie et al. 1991).

In particular, the Mars Express (MEX) mission, in Martian orbit since 2003, has improved our knowledge of the interior of Phobos. The determination of its bulk density has been significantly improved (Andert et al. 2010; Willner et al. 2010), and has been interpreted as evidence supporting in situ formation for this moon (Andert et al. 2010; Rosenblatt et al. 2010). The Mars Express spacecraft has also provided new observations of Phobos' surface from its spectral remote-sensing instruments. A silicate composition for Phobos has been inferred from some of these data, which have been interpreted as additional evidence in favor of in situ formation for Phobos (Giuranna et al. 2011), but other data are in agreement with previous remote-sensing observations (Gondet et al. 2008), emphasizing the ambiguity on the determination of Phobos' surface composition from remote-sensing data (Pieters 2010).

The goal of this review is to emphasize the importance of the internal structure of the Martian moons on our understanding of their origin. The pros and cons of the scenarios of origin proposed in the literature will be reviewed by putting forward the link with the interior of the moons, and by showing how a better knowledge of this internal structure may help to solve the challenges raised by these scenarios. Some perspectives will also be given since the still ongoing Mars Express mission and the soon-coming Phobos-Grunt (or Phobos-Soil) mission (due for launch in November 2011) will offer unique opportunities to refine our understanding of the interior of Phobos, and perhaps to provide a definitive answer about the origin of the Martian moons.

2 The scenarios of origin of the Martian moons

Two kinds of scenario have been proposed to explain the origin of Phobos and Deimos: the capture scenario and the in situ formation scenario (in Mars' orbit). These scenarios have been developed based on the observations of the American Mariner-9, Viking-1 and Viking-2 and the former Soviet Union Phobos-2 spacecraft.

2.1 The capture scenario

The morphological characteristics of Phobos and Deimos In contrast to the Earth's moon, the Martian moons are small and irregularly shaped bodies (with a diameter not larger than about 20 km, see Table 1 and Fig. 1). Their surface is very dark with an albedo not larger than a few percent. These morphological characteristics of Phobos and Deimos are very similar to those of numerous low-albedo and small-sized asteroids. The surface of Phobos shows numerous grooves, which are also observed on some asteroids. Several authors have claimed that Phobos' grooves might have been formed during the capture by Mars (e.g. Hunten 1979; Pollack et al. 1979). However, other studies have proposed a different origin for these grooves, which do not necessarily imply a capture process; for example, chains of secondary impact craters associated with the formation of the Stickney crater, (see Hamelin 2011 for a short review), or ejecta from impacts on Mars (Murray et al. 2006). These grooves are not observed on Deimos' surface.

Similarly to the asteroids, the surface of Phobos is highly cratered. This suggests that Phobos' surface is at least 1 billion years old (Pollack 1977; Thomas and Veverka 1980). This age estimation is based on the crater count technique, which assumes that

Table 1 Shape (best-fit ellipsoid), volume, mass and density of Phobos and Deimos, from (1) Willner et al. (2010), (2) Rosenblatt et al. (2008), (3) Thomas (1993), (4) Jacobson (2010), (5) This study

	Phobos	Deimos
Radius (in km)	$13.0 \times 11.39 \times 9.07$ (1)	$7.5 \times 6.1 \times 5.2$ (3)
Volume (in km^3)	$5748 +/- 190$ (1)	$1017 +/- 130$ (3)
Mass (in 10^{16} kg)	$1.06 +/- 0.03$ (2)	$0.151 +/- 0.003$ (4)
Density (in g/cm^3)	$1.85 +/- 0.07$ (5)	$1.48 +/- 0.22$ (5)

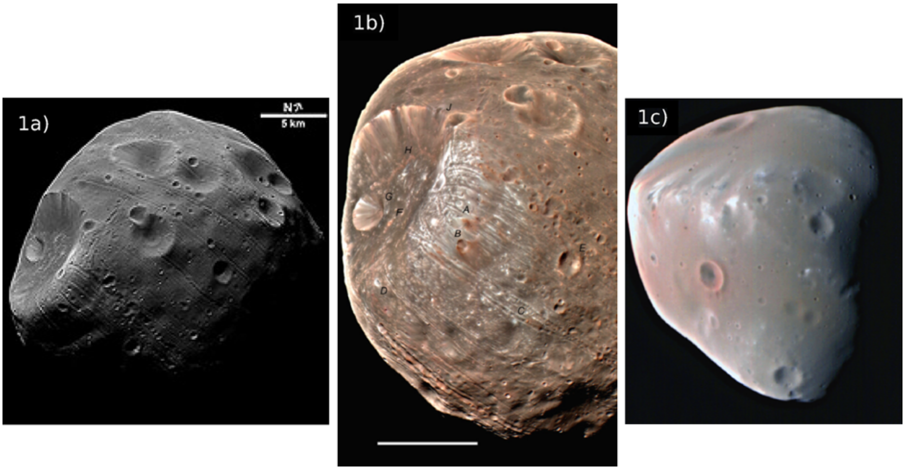

Fig. 1 Recent images of the Martian moons from current Mars orbiting spacecraft. (**1a**) Phobos from Mars Express High Stereoscopic Resolution Camera (courtesy DLR/ESA); (**1b**) Phobos and (**1c**) Deimos from Mars Reconnaissance Orbiter High-Resolution Imaging Science Experiment (Thomas et al. 2010)

all the identified craters result from a bombardment of the surface with the same rate as for the Lunar surface (Lambeck 1979). Phobos' surface shows, however, a large crater (the Stickney crater with a diameter of 10 km, see Fig. 1), whose formation was likely associated with numerous ejecta that may have formed numerous secondary craters. As a consequence, the age of Phobos' surface would be incorrectly estimated using the crater count technique. The craters on Deimos' surface appear more subdued than on Phobos probably due to a thicker regolith on Deimos than on Phobos (Thomas et al. 1992).

The morphological similarities between Phobos and Deimos, on the one hand, and numerous asteroids, on the other, have incited scientists to propose that both moons are asteroids from the main belt located between Mars and Jupiter subsequently captured, under suitable conditions, by Mars' gravitational attraction. This idea has been re-inforced by the assessment of the surface composition of both moons from the remote-sensing observations of their surfaces in the Visible-Near-infraRed (ViS-NiR) wavelength band (about from 0.4 µm to 4 µm).

The composition of the surface of Phobos and Deimos The first reflectance spectra of Phobos measured by the Viking-1 spacecraft were found to be similar to those of low-albedo carbonaceous C-type asteroids of the main belt (Pang et al. 1978;

Pollack et al. 1978). However, these spectra were obtained with a coarse spatial resolution and represent spectral properties of the surface averaged over the disk of Phobos (Murchie and Erard 1996). The spectra from the Phobos-2 mission for Phobos (Murchie et al. 1991) and from the Hubble Space Telescope for both Phobos and Deimos (Rivkin et al. 2002), which have better spatial and spectral resolution, were found to better match those of low-albedo carbonaceous D or T-type asteroids of the outer belt and of the Trojans at Jupiter orbit (at a distance of 5 AU from the Sun). However, no absorption bands, suitable as diagnostics of the surface composition, are visible on the moons' spectra at the resolution of these instruments. These spectra rather show a spectral slope toward near-infrared wavelengths corresponding to a reddening of the surface color. The spectra of the trailing hemisphere of Phobos have a more pronounced reddening than those of the leading hemisphere of Phobos. The trailing and leading hemispheres correspond to the so-called 'red' and 'blue' units of Phobos' surface, respectively. The spectra of the 'red' and 'blue' units best match those of D-type and T-type asteroids, respectively (Murchie and Erard 1996; Rivkin et al. 2002). However, it is not clear whether these two color units really are caused by compositional variations on Phobos or spatial variations of its physical surface properties (Murchie and Erard 1996; Pieters 2010). The spectra of Deimos are similar to Phobos' red unit ones, and they do not show any spatial variations.

This similarity between the surface spectra of the Martian moons and of the D or T-type asteroids suggests that the two moons are composed of the same material as these asteroids, i.e. carbonaceous material. This material is expected to have condensed in the solar nebula at a distance far away from Mars and then moved inwards the inner solar system to account for the carbonaceous composition of Phobos and Deimos. Recent dynamical models of early solar system dynamics (namely the *Nice Model*, Tsiganis et al. 2005; Gomes et al. 2005) supports the idea of migration of material from the outer region toward the inner region of the solar system, thus giving additional credentials to the capture scenario to account for the formation of the two moons of Mars.

This scenario is, however, weakened by some ambiguities in the interpretation of the reflectance spectra and by the difficulty to account for all the observations of the Martian moons, such as their current orbits around Mars.

2.2 The challenges raised by the capture scenario

2.2.1 The ambiguity on the surface composition

No matching between Phobos and Deimos spectra and those of low-albedo carbonaceous meteoritic samples, recognized as the material analog of low-albedo carbonaceous asteroids, has been found so far (e.g. Murchie et al. 1991; Murchie and Erard 1996; Vernazza et al. 2010). Indeed, the reddened slope of the spectra of the class CI/CM carbonaceous chondrite (like the Murchison meteorite sample) has been found to fit the slope of the Phobos blue unit spectrum (Rivkin et al. 2002), but in contrast to this meteorite spectrum, Phobos' spectrum does not show the complex absorption band at 3 μm (Bibring et al. 1989). This absorption band is the signature of hydration of the material, and its absence in Phobos and Deimos spectra has been interpreted either as non-hydration of the surface or as a dehydration process having occurred at the surface of both moons. Rivkin et al. (2002) proposed a heating metamorphism to account for dehydration, but without providing the origin of this process. The Tagish Lake carbonaceous meteorite has been recognized as one possible material analog to D-type asteroids (Brown et al. 2000; Hiroi et al. 2001). The albedo of this meteorite is as dark as the Phobos and Deimos albedo and fits very well the spectrum of the Phobos blue unit (Vernazza et al. 2010), except for the presence of the 3 μm absorption band in the Tagish Lake spectrum (Hiroi et al. 2001). The spectra of class CO/CV/CR anhydrous carbonaceous chondrites do not show the 3 μm band, but they do not match the reddened shape of the moon spectra and their albedo is significantly higher than the albedo of Phobos and Deimos (Murchie and Erard 1996).

In order to explain the partial matching between the moons' spectra and carbonaceous meteorite spectra, it has been proposed that the spectra of the moons' surfaces may have been altered by the space weathering effect (Murchie et al. 1991; Murchie and Erard 1996; Rivkin et al. 2002). Indeed, the charged particles of the solar wind and the micrometeoritic bombardment on airless body surfaces are known to subdue the absorption bands and to redden the ViS-NiR spectra of these surfaces (Clark et al. 2002). As a consequence, the 'true' composition of the airless bodies may be hidden in these remote-sensing spectra, especially when trying to detect this composition by spectral matching (Gaffey 2010) as done for Phobos and Deimos. This effect of space exposure weatherings has been studied in detail for the surface of the Moon thanks to the Lunar samples returned to Earth. Although no asteroid or Martian moon samples are available yet, simulations of this process on some meteoritic samples have been performed on the basis of Lunar studies. The simulated weathered spectra of the Mighei CM class hydrated carbonaceous chondrites have revealed that the 3 μm band can be significantly subdued[1] (Moroz et al. 2004). In turn, this suggests that the non-detection of this absorption in the spectra of Phobos and Deimos does not necessarily mean that their surfaces are not composed of hydrated minerals. However, the reddened slope of both Phobos blue and red unit spectra could not be reproduced by these simulated weathered spectra (Moroz et al. 2004). Recently performed simulations of the space weathering effect on a Tagish Lake meteorite sample have revealed that the slope of the Phobos red unit and Deimos spectra could not be reproduced (Vernazza et al. 2010). Some authors have argued that the Kaidun meteorite (an unusual carbonaceous meteorite, e.g. Ivanov and Zolensky 2003; Ivanov 2004) might have Phobos as its parent-body,[2] but no spectral comparison with Phobos surface spectra has been made yet in order to further investigate this interpretation.

The ordinary chondrites of silicate composition are much brighter than the Phobos and Deimos surface, but some of them, the black chondrites, are nearly as dark as the Martian moons' surfaces. These black chondrites have nearly featureless spectra

[1] Moroz et al. (2004) have used irradiation of meteoritic samples with a microsecond pulsed laser in order to simulate the effect of the micrometeoritic bombardment on airless body surfaces. They found that this effect may cause a dehydration of the surface, yielding an attenuation of the 3 μm absorption band.

[2] These authors studied the composition of the Kaidun meteorite and found unusual minor components corresponding to deeply differentiated rock (like alkaline-rich rock). They argued that the parent-body of this meteorite is a carbonaceous chondrite satellite of a large differentiated planet, and proposed Phobos as a good candidate.

Table 2 Current orbit of Phobos and Deimos from Jacobson (2010)

	Phobos	Deimos
Semi-major axis (in km)	9375.0	23458.0
Eccentricity	0.01511	0.00024
Inclination to Mars' equator (in °)	1.0756	1.7878
Period of revolution (in hours, minutes, seconds)	7h 39′ 19.47″	30h 18′ 1.36″

with a spectral slope flatter than the spectral slope of the Martian moons (Britt and Pieters 1989). Murchie et al. (1991) and Murchie and Erard (1996) have quoted that a highly space weathered version of this material might be considered as a spectral analog to Phobos material (although no simulations of space weathering on such meteoritic sample have been performed yet). A silicate composition is also in agreement with the detection of a tiny absorption band around 1 μm in Phobos spectra, suggesting the presence of olivine and pyroxene minerals (Gendrin et al. 2005). Therefore, the surface of Phobos may be composed of highly weathered (or matured) silicate material instead of carbonaceous material although this interpretation requires a significant space weathering effect or a high maturity of Phobos' soil (even larger than the maturity of the Lunar soil, Murchie and Erard 1996).

Alternative interpretations of the ViS-NiR reflectance spectra would be that analogs to Phobos and Deimos material are not in the current meteoritic collection (thus unknown)[3] or that the surface of the two moons would be recovered by a fine layer of 'alien' material (Pieters 2010). These latter interpretations emphasize the ambiguity of a determination of the material composing the two moons from remote-sensing observations of their surfaces.

2.2.2 The past evolution of the orbits

As soon as the Martian moons were discovered, Earth-based telescopic observations have been used to determine their current orbits around Mars. Both orbits are near-circular and near-equatorial at a distance to the center of Mars of about 2.76 R_M and 6.92 R_M for Phobos and Deimos, respectively (Burns 1992, see also Table 2. R_M is the radius of Mars (3396 km)). As first quoted by Sharpless (1945), Phobos shows a secular acceleration along its orbit, compared to its Keplerian rate, making it slowly spiraling in toward Mars. The estimation of the orbital acceleration of Phobos has since been refined from fits of numerous orbital models to astrometric observations made both from Earth and by spacecraft (e.g. Sinclair 1989; Lainey et al. 2007; Shishov 2008; Jacobson 2010). The most recent estimates of the secular acceleration is 1.27×10^{-3} deg/yr^2 (Lainey et al. 2007; Jacobson 2010), corresponding to an orbital decay rate of about 20 cm/yr. This secular acceleration has been explained as resulting from the solid-body tides raised by Phobos in Mars (e.g. Burns 1992).[4]

[3] More material could be gathered by the space missions aiming to bring samples of dark asteroids back to Earth, such as the Hayabusa-2 or Marco Polo missions.

[4] The tidal bulge raised by Phobos in Mars is indeed shifted with respect to the Mars-Phobos direction. This shift is due to the fact that Mars does not instantaneously deform under the gravitational stress exerted by

The same mechanism is expected to recede Deimos' orbit away from Mars (as for the Earth's Moon) because Deimos orbit is beyond the synchronous orbit, which lies at 6 R_M. The predicted rate of the orbit deceleration of Deimos is, however, less than 1 percent of Phobos' rate (Burns 1992), which is too small to have been measured so far.

The models of tidal orbital evolution have also been used to test the capture scenario by integrating the orbital changes backward in time (back to 4.6 Ga, e.g. Burns 1992). Indeed, in the capture scenario, the initial orbit of a captured asteroid is expected to be highly elliptical and in a near-heliocentric plane (where most of the asteroids orbit the Sun) while Phobos and Deimos orbits are near-circular and near-equatorial. Therefore, the capture scenario must imply that processes after capture have circularized and changed the inclination of the orbit of the captured asteroid.

The efficiency of tidal orbital changes depends, however, on the tidal dissipation rate in Mars as well as in the moons (Lambeck 1979). The tidal dissipation rate inside the moons can play a major role since it significantly accelerates the orbital changes particularly when the orbit is highly eccentric (Lambeck 1979); the larger the dissipation rate is in the moon, the faster the orbit changes. In the case of Phobos the rate of eccentricity changes required by the capture scenario can be achieved over the last 4.6 Ga, but given a high tidal dissipation rate in Phobos or a tidal quality factor Q about 5 times lower than that of Mars (assuming a bulk rigidity of Phobos of 0.2 GPa Lambeck 1979). This low tidal quality factor seems, however, difficult to reconcile with a rocky monolithic Phobos (see Sect. 2.2.3). The inclination changes require an even lower Q factor of Phobos (up to 25 times lower than that of Mars, Mignard 1981[5]), which is more relevant to icy material than to rocky material (see Sect. 2.2.3). Moreover, this tidal orbital evolution scenario implies that Phobos' orbit would have crossed Deimos' orbit, thus making the collision between both bodies likely since the collision timescales (about 10^5 years, Cazenave et al. 1981) are much shorter than the tidal timescales (some 10^9 years, Burns 1992). In the case of Deimos, the rate of orbital changes is, however, too slow to account for this moon's current orbit by tidal orbital evolution over the last 4.6 Gyr (Lambeck 1979; Szeto 1983), and does not permit a reconciliation with the capture scenario (using tidally induced orbital changes).

The difficulty to change Phobos' orbit inclination, from the ecliptic plane to the Martian equatorial plane, has been used to argue against the capture scenario for this body (Burns 1992). Nevertheless, the problem of inclination changes might be overcome if one assumes either that both moons have been captured in the current equatorial plane of Mars or that Mars' equatorial plane was in the ecliptic plane at the time of capture. The assumption of a capture in the equatorial plane of Mars requires that the asteroids would have had an initial orbital plane inclination off the ecliptic plane of about 20°, which may not be unexpected especially for outer-belt

Phobos. As Phobos is orbiting at a rate faster than Mars' spin rate, the tidal bulge lags behind with respect to Phobos' position as seen from Mars. In turn, a gravitational torque is exerted on Phobos, which allows for the transfer of orbital energy to Mars rotational energy, causing the orbital decay of Phobos and the acceleration of Mars' spin rate.

[5] Assuming the same rigidity of 0.2 GPa for Phobos as in Lambeck (1979).

asteroids (Lambeck 1979). The assumption of a Mars' equatorial plane in the ecliptic plane at the time of capture is not unrealistic as shown by recent models of Mars' obliquity variations (Laskar and Robutel 1993). However, these variations are faster than the tidal orbital changes, and an asteroid captured following this scenario, must have an orbit after capture bounded within roughly 13 R_M in order for its orbital plane to remain in the equatorial plane of Mars.[6]

In order to help the circularization of the initial elliptical orbit of a captured asteroid, another process has been proposed, namely the drag effect in the planetary nebula surrounding the proto-Mars (e.g. Kilgore et al. 1978; Sasaki 1990). This process is especially interesting for Deimos for which tidal effects are too small to change its orbit over the last 4.6 Ga (Burns 1992). The drag process is particularly efficient in circularizing an orbit, but it requires a very early capture and a suitable density of the planetary nebula in order to prevent the captured asteroid from crashing on Mars (Sasaki 1990). In addition, the gas nebula has to extend, at least, beyond the current orbit of Deimos and its lifetime has to be shorter than the Phobos–Deimos collision lifetime (10^5 years). In addition, the drag-induced circularized orbits have to be at a distance of below and beyond the synchronous orbit (at 6 R_M) for Phobos and Deimos, respectively, to be compatible with its following tidal decay over the last 4.6 Gyr (Burns 1992). In view of these restrictive conditions on the gas nebula, the drag effect appears to be an ad hoc process to account for the current orbits of the two moons in the capture scenario.

The numerous studies on the past evolution of the orbit of Phobos and Deimos have shown that orbital changes, required by the capture scenario, cannot easily be explained. In particular, the tidally induced orbital changes require a too low tidal quality factor for rocky monolithic Phobos and Deimos (see Sect. 2.2.3). This difficulty of finding an efficient mechanism of orbital evolution has been used to argue against the capture scenario. Nevertheless, the same difficulty can be reversed in favor of the capture scenario. Indeed, such difficulties could explain why only a few asteroids might have been captured by Mars with regard to the huge amount of potential candidates in the asteroid population between Mars and Jupiter.

2.2.3 The interior of the moons and their orbital evolution

As shown in previous studies, the past evolution of the orbits of Phobos and Deimos depends on the tides raised by the moons in Mars and the tides raised by Mars in the moons. The rate of these tidally induced orbital changes depend on the ratio of the dissipative properties in the moons and in Mars by the following relationship (from Mignard 1981):

$$A = \frac{k_2'/Q'}{k_2/Q}\left(\frac{M}{M'}\right)^2\left(\frac{R'}{R}\right)^5 \tag{1}$$

where k_2 is the tidal Love number, Q the tidal quality factor, M the mass and R the radius of Mars. The same quantities with ($'$) stand for the moons. The k_2 number describes the ability of a body to deform under the tidal stresses. It is roughly inversely

[6] Beyond 13 R_M, the third-body gravitational attraction of the Sun is large enough to maintain the orbit of a captured asteroid in the ecliptic plane (Burns 1992; Mignard 1981).

proportional to the rigidity (or shear modulus) of the body (Munk and MacDonald 1960). The tidal quality factor Q is inversely proportional to the fraction of the deformation energy dissipated in the body per tidal cycle (the smaller the Q factor, the larger the dissipation rate).

In his study, Lambeck (1979) has assumed that the Q factor in Phobos is about 5 times lower than in Mars,[7] and a relatively low rigidity of 0.2 GPa for a Phobos composed of hydrated carbonaceous chondritic material. Unfortunately, no measurements of either rigidity or Q factor of the carbonaceous meteorite samples are available at tidal frequencies, and the validity of the assumptions of Lambeck (1979) cannot be verified. However, several authors presume an upper bound of 100 for the Q factor of carbonaceous chondritic material (Efroimsky and Lazarian 2000). If we consider this upper bound as a more realistic Q value for carbonaceous material, then the dissipation rate would be ten times lower than assumed by Lambeck (1979). In addition, the rigidity assumed by Lambeck (1979) is about 20 times lower than the rigidity measured for some samples of ordinary chondrites. As these chondrites are recognized as the material analog of S-type asteroids, the computation of Lambeck (1979) also shows that it would be difficult to achieve the orbit eccentricity changes required by the capture of such kind of asteroids.

On the other hand, Yoder (1982) proposed a lower bound for the tidal rigidity of Phobos of 1 GPa (for a Q value of 100), on the basis of a model of orbital evolution mainly driven by series of gravitational resonances. However, his estimate is based on the assumption that the orbital eccentricity of Phobos was zero 1 Ga ago. Under that assumption, a high dissipation rate inside Phobos is not required in order to account for the current eccentricity of its orbit (see Table 2).

Mignard (1981) has shown that the inclination variations required by the capture scenario might be achieved for a high tidal dissipation rate in the moon (for a value of the A factor of up to 100), yielding Q' as low as 2 (taking M, R, k_2/Q, R', M' and k' values as in Lambeck 1979). Such a low value of Q is more relevant to icy materials than to rocky material (McCarthy and Castillo-Rogez 2011). In turn, it shows that significant tidally induced orbital changes, as required by the capture scenario, cannot be achieved for a monolithic rocky Phobos. However, if Phobos is not monolithic but made of rocky material and additional material(s) with high dissipative properties, its bulk dissipative properties may be significantly enhanced with respect to the rock monolithic case. That point will be discussed in Sect. 4.

It is also important to note that the values of k_2 and Q depend on the tidal frequency (e.g. Bills et al. 2005), i.e. on the period of revolution (or on the semi-major axis) of the moon around Mars. Some previous studies have taken into account the Q dependency on tidal frequencies in their computation of orbital evolution (e.g. Lambeck 1979; Mignard 1981). However, recent works have shown that the tidal-frequency scaling laws used for Q in these studies were not suitable for Mars' rheology (Efroimsky and Lainey 2007). These authors have shown that a larger dissipation rate in Mars is expected from more realistic Q-scaling laws, hence a faster orbital evolution of Phobos than predicted by previous studies. However, these new

[7] Lambeck (1979) considered a Q value of 50 for Mars at the present tidal frequency raised by Phobos. The current value is estimated around 80 (Lainey et al. 2007; Jacobson 2010).

Q-scaling laws have been applied to the future and not the past evolution of the orbit of Phobos (Efroimsky and Lainey 2007). Further studies are needed to assess how much the previous results of the past evolution of Phobos' orbit could be changed using these more realistic Q-scaling laws.

2.3 The in situ formation scenarios

Several authors have proposed scenarios of in situ formation in orbit around Mars in order to overcome the challenges raised by the capture scenario. One of these in situ scenarios (Singer 2003, 2007) proposes that Phobos and Deimos are the last two remnants of a parent larger moon, which has been captured by Mars and then destroyed when entering into the Roche limit. The orbit of a captured large body is easier to change so that it could orbit in the equatorial plane of Mars before being destroyed by tidal forces. Therefore, the remnants are expected to orbit Mars in the equatorial plane, which is consistent with the current Martian moons' orbit. In this scenario the orbit of most of the remnants are expected to tidally spiral toward Mars, and so to crash onto it. The resulting impact craters may be consistent with the distribution of the oblique impact craters at Mars' surface (Schultz and Lutz-Garihan 1982). This scenario is not inconsistent with a carbonaceous composition of Phobos and Deimos, if one assumes that the early larger moon had the same composition. However, this scenario requires that some remnants orbited Mars just beyond the synchronous orbit (6 R_M) in order to account for the current position of Deimos, which is difficult to reconcile with the disruption of this early moon below its Roche limit (about 2.5 R_M).[8]

An alternative scenario proposes that Phobos and Deimos have been formed from the re-accretion of debris blasted into Mars' orbit by the collision between Mars and a large body with a diameter of about 1800 km (Craddock 1994, 2011). In this scenario, the debris inserted into orbit formed a disk around Mars from which moonlets accreted, Phobos and Deimos being the last two remnants of this moonlet population (the other moonlets would have crashed onto Mars due to the tidal decay of their orbits, Craddock 2011). In addition, the orbit of the moonlets are expected to be near-equatorial and near-circular, and a carbonaceous composition for Phobos and Deimos can be accounted for, if the large body impactor was of carbonaceous composition.[9] However, this scenario has not been studied in detail yet, and leaves unexplained several observations; for example, it requires that the accretional disk could extend beyond the synchronous orbit in order to account for the formation of Deimos.

Other authors have proposed that Phobos and Deimos could have been formed from a debris disk left over from the formation of Mars (i.e. co-accreted with Mars, Safronov et al. 1986). This scenario permits to account for the current near-equatorial and near-circular orbits of both moons, but it implies that Phobos and Deimos have the composition of the building blocks that formed Mars, which does not seem obvious to reconcile with a carbonaceous composition.

[8] The tidal effects on the orbit of the close-to-Mars remnants tend to make them spiraling toward Mars and not to recede them away from Mars.

[9] The debris blasted into Mars' orbit from such a collision can come mainly from the impactor, depending on the velocity and angle of the collision, as shown in the Earth–Moon case (e.g. Cameron 1986).

None of these in situ formation scenarios has been studied in detail, they may therefore be considered to be ad hoc scenarios. Nevertheless, observations do not strongly argue against them. In particular, the possible carbonaceous composition for Phobos and Deimos can be accommodated, and therefore this composition alone does not preclude an in situ formation scenario for the two moons.

3 Recent Mars Express observations of Phobos surface and density properties

The Mars Express (MEX) spacecraft has been orbiting Mars since the end of 2003. This mission, initially designed to study the surface, interior and atmosphere of Mars, has also been used to study the surface and interior of Phobos. Thanks to its eccentric orbit, MEX is the only spacecraft, among the current Mars orbiting spacecraft, able to perform close flybys of Phobos. To date about 120 flybys have been performed at a distance closer than 1000 km (the closest was 77 km on March 2010). These flybys have permitted to improving the observations of previous missions, on the one hand, and to perform new kinds of observation, on the other hand. The first results of these MEX observations, as well as some observations from the Mars Global Surveyor (MGS) and Mars Reconnaissance Orbiter (MRO) spacecraft, relevant to the origin of the moons, are summarized in this section.

3.1 The composition of the surface of Phobos

New ViS/NiR reflectance spectra have been obtained by the OMEGA[10] instrument onboard MEX for the surface of Phobos (Gondet et al. 2008) and by the CRISM[11] instrument onboard MRO for the surface of Phobos and Deimos (Murchie et al. 2008). These spectra have confirmed the previous observations (i.e. reddening of the spectra and lack of absorption bands at the threshold of detectability of the instruments[12]). The new high-spatial-resolution images from MEX/HSRC[13] and from MRO/HiRiSE[14] (see Fig. 1) confirmed the previous observations obtained at lower resolution, but have also revealed details which indicate a far more complex relationship between the two color units (‘blue’ and ‘red’) on Phobos and also smaller-scale color variations on Deimos (Thomas et al. 2010). This complexity at different spatial scales still leaves open the question raised by the previous data on Phobos: Are the color variations due to compositional variations or to different degrees of weathering (or anything else) (Pieters 2010)?

Some emissivity spectra of the surface of Phobos have also been acquired by the Planetary Fourier Spectrometer (PFS) onboard MEX in the Infra-Red wavelength

[10]Observatoire pour la Minéralogie, l’Eau, les Glaces et l’Activité.

[11]Compact Reconnaissance Imaging Spectrometer for Mars.

[12]Murchie et al. (2008) argue that a signature of carbonaceous material has been detected on the CRISM spectra of both Phobos red unit and Deimos, but such a signature has not been detected on the OMEGA spectra of Phobos (Gondet et al. 2008).

[13]High Stereoscopic Resolution Camera.

[14]High-Resolution Imaging Science Experiment.

domain (from about 5 μm to 50 μm). Along with emissivity spectra by the Thermal Emissivity Spectrometer (TES) onboard MGS (Roush and Hogan 2000; Palomba et al. 2005, 2010), these MEX emissivity spectra have been used to infer the composition of Phobos' surface (Giuranna et al. 2011). Surprisingly, these spectra do not match those of carbonaceous material, and are rather consistent with silicate material. This discrepancy of compositional signature between Infra-Red emissivity and ViS/NiR reflectance spectra may be explained by the fact that space weathering affects spectra at different wavelengths differently (Giuranna et al. 2011). However, this effect has been poorly documented, especially for emissivity spectra in Phobos' surface conditions, therefore more laboratory experiments are needed to confirm or disprove this explanation.

This interpretation of the emissivity spectra raises a pertinent point about the origin of Phobos. Indeed, if Phobos is made of silicate material, it means that it is not required anymore to bring material formed in the outer solar system into Mars' orbit, which is the main argument supporting the capture scenario. However, this scenario cannot be precluded on the basis of the emissivity spectra alone, since they could also be interpreted as resulting from achondrite silicate material (Howardite, Giuranna et al. 2011); this would be consistent with the capture of an asteroid composed of such material.

The new remote-sensing observations of the surface of Phobos and Deimos by MEX, MRO and MGS do not solve the ambiguity of the composition of these surfaces. The material composing the Martian moons could be either carbonaceous or silicate or of another kind not available in the meteoritic collection.

3.2 The density of Phobos and Deimos

Measurements by the radio-science experiment (MaRS) on Mars Express have significantly improved the mass estimate of Phobos (Rosenblatt et al. 2008; Andert et al. 2010). Both the elliptical orbit of MEX[15] and the improved ephemeris of Phobos derived from new astrometric data of the Super Resolution Camera (SRC) onboard MEX (Lainey et al. 2007) contributed to this refinement. The images of the HRSC/SRC cameras have also permitted to better determine the volume of Phobos (Willner et al. 2010). These new data have been used to determine the bulk density of Phobos as $1.85 + / - 0.07\ \mathrm{g/cm^3}$ (with a relative error of only 4%, see Table 1).[16] More recently, the mass of Deimos has also been redetermined by a re-analysis of the tracking data of the Viking-2 close flyby (closest approach to 30 km), using new Deimos ephemeris, improved with the MRO images (Jacobson 2010). The density of Deimos is now estimated to be $1.48 + / - 0.22\ \mathrm{g/cm^3}$ (see Table 1). Recent observations of the surfaces of Phobos and Deimos by the Arecibo radar facilities have led to

[15] Indeed, MEX orbit is more sensitive to Phobos' third-body acceleration than spacecraft orbiting closer to Mars. It has permitted to estimate Phobos' mass using two different approaches: the first consists of accumulating radio-tracking data over a long period of time (without close flybys, Rosenblatt et al. 2008) and the second consists of using radio-tracking data acquired just during flybys (Andert et al. 2010). Both approaches have given consistent solutions for the mass of Phobos.

[16] Previous estimates, from the Viking and Phobos-2 mass solutions and Viking volume solutions were in the range 1.57 to $2.20\ \mathrm{g/cm^3}$, see Smith et al. (1995).

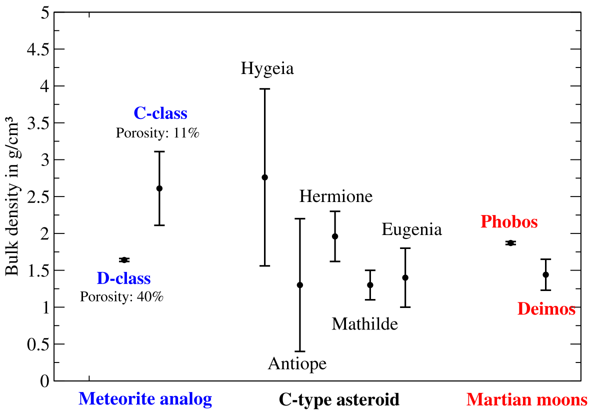

Fig. 2 Bulk density of Phobos and Deimos (see Table 1), of low-albedo carbonaceous asteroids and of carbonaceous meteoritic samples (asteroid and meteorite densities are from Britt et al. 2002)

an upper bound of the bulk density of the first tens centimeters of the surface layer of $1.6 +/- 0.3\ \mathrm{g/cm^3}$ and $1.1 +/- 0.3\ \mathrm{g/cm^3}$ for Phobos and Deimos, respectively (Busch et al. 2007). These surface densities are lower than the bulk density of the two moons by up to about 30% and 46% for Phobos and Deimos, respectively. This could be interpreted as porosity of the soil of the moons resulting from the micrometeoritic bombardment of their surfaces (i.e. regolith formation, Busch et al. 2007).

The density of both moons is similar to the density of low-albedo carbonaceous C-type asteroids, although it would be in the lower part of the density range of these asteroids (Fig. 2). Only a few data on the bulk density of D-type asteroids are available. The Jupiter Trojan object 617 Patroclus has a density of $0.8 +/- 0.2\ \mathrm{g/cm^3}$ (Marchis et al. 2006); this is much lower than the density of Phobos and Deimos.[17] Although the measurement of the bulk density of asteroids is generally difficult, the available data indicate that the densities are generally lower than the density of their carbonaceous meteoritic analog (Fig. 2). A significant amount of porosity in their interior is thought to account for their low densities (e.g. Britt et al. 2002; Consolmagno et al. 2008). The Martian moons also have a bulk density lower than most of the samples of carbonaceous material (Fig. 2); this requires porosity (i.e. voids) and/or light elements like water ice in their interiors (cf. Avanesov et al. 1991; Murchie et al. 1991).

4 The internal structure and origin of the Martian moons

To date, the question about the origin of Phobos and Deimos was discussed only based on observations concerning their surface and their orbit. The improved determination of their bulk density now allows us to include their bulk internal structure (i.e. possible content of porosity and water ice in their interior) in the discussion about their origin.

[17] The density of the 624 Hektor Trojan object has been measured as $2.5\ \mathrm{g/cm^3}$, but with a large uncertainty of 45% (Lacerda and Jewitt 2007), which encompasses the density of Phobos and Deimos.

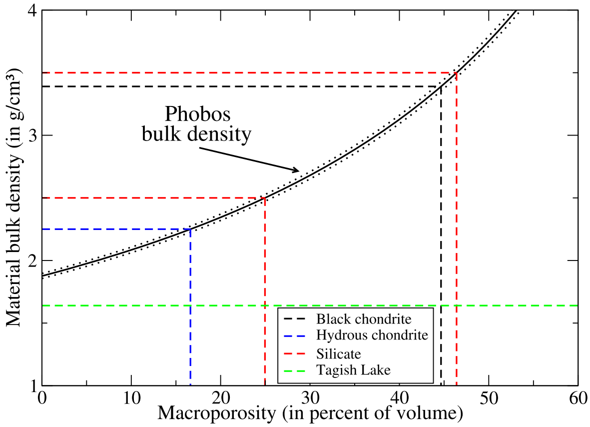

Fig. 3 Estimation of macroporosity inside Phobos that fits its bulk density (*black solid curves*) within its error bar (black dotted curves), given the bulk density of a range of rocky material analogs composing the moon (see Table 3)

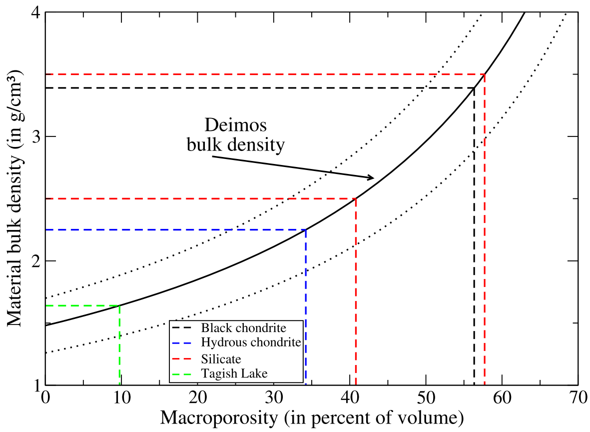

Fig. 4 Same as for Fig. 3 but for Deimos

Table 3 Bulk density of analog to Martian moon material, considered in this study. (1) Consolmagno et al. (2008), (2) Hildebrand et al. (2006), (3) Average value from Britt and Consolmagno (2008)

Material analog	Bulk density (in g/cm^3)
Hydrated carbonaceous chondrite	2.25 (1)
Tagish Lake	1.64 (2)
Black chondrite	3.39 (3)
Silicate (lower bound)	2.50
Silicate (upper bound)	3.50

4.1 Porosity inside Phobos and Deimos

The remote-sensing data cannot determine the composition of the bulk moon since they only sense the first microns of the surface layer. Impact craters may provide a view of the interior by deeper layers exposed at the surface.[18] The color variations

[18] Deeper layers are expected to be exposed at the surface at the central peak and at the rim of the craters.

revealed by the MRO images show a complex relationship between the morphological structures at the surface of the moons (Thomas et al. 2010; Pieters 2010), but do not allow establishing if they really reveal compositional variations between the surface and the interior. For simplicity, we will therefore assume in the following that the rocky compound of the interior of the moons corresponds to one single rock material. If only porosity and no light element inside each moon is considered in addition to the rocky compound with density ρ_a, the porosity content Φ needed to fit the observed bulk density ρ_b within its error bars is given by

$$\Phi = 1 - \frac{\rho_b}{\rho_a} \tag{2}$$

Figures 3 and 4 show the range of porosity inside Phobos and Deimos, respectively, computed for a large range of plausible material analogs. A porosity of about 15% is obtained for Phobos, assuming a hydrated carbonaceous material analog. This is lower than the 20% to 60% range inferred for the C-type asteroids made of such material (Britt et al. 2002). Silicate material analogs have larger bulk density, and the porosity estimate is larger (up to 45% of the volume for black chondrite or dense silicate material analogs, see Fig. 3). This high porosity is significantly larger than the range estimated for S-type asteroids (20% or less, except for Near-Earth S-type asteroids which have macroporosity estimated at about 40%, Consolmagno et al. 2008). The Tagish Lake meteorite material has been identified as an analog to D-type asteroid material (Hiroi et al. 2001), and it has a bulk density of 1.67 $\mathrm{g/cm^3}$ $+/-$ 0.02 $\mathrm{g/cm^3}$ (Hildebrand et al. 2006). This density is lower than Phobos' bulk density by at least 0.12 $\mathrm{g/cm^3}$ (or 7.3 percent of the Tagish Lake material density). A compression of this material is therefore required for achieving the Phobos bulk density value. Phobos is, however, too small to compress by self-gravity the Tagish Lake material by such an amount. Impacts may also increase the bulk density by reducing the porosity inside a Phobos composed of the Tagish Lake material (this material has indeed a grain porosity, or microporosity, of 40%, Hildebrand et al. 2006). However, in such a case, the energy of impacts is confined to a small volume close to the impact site and cannot propagate through the entire volume of the body (Richardson et al. 2002). This makes it difficult to achieve a significant increase of the bulk density. Therefore, the larger density of Phobos suggests that the origin of Phobos is inconsistent with the capture of a D-type asteroid. However, the density of the D-type material relies on a single meteorite sample that may not be representative of the entire D-type population. Indeed, the density of the material condensed in the solar nebula at a heliocentric distance of about 5 AU is presumed to be between 1.5 and 2 $\mathrm{g/cm^3}$ (Consolmagno et al. 2008). Therefore, if Phobos is an object condensed at such large distances, its bulk density indicates that it would contain between 0% and 20% of macroporosity (Fig. 3). This is in contradiction with the available macroporosity estimates inside solar system objects, which show values generally larger than 40% at distances beyond 3 AU (Consolmagno et al. 2008). The same contradiction is also raised from the comparison of the macroporosity estimates of Phobos and of C and S-type asteroids as quoted above. As the general trend of macroporosity in the solar system seems to reflect the accretional and collisional environment of the early solar system (Consolmagno et al. 2008), this contradiction may indicate that Phobos would have formed in a different environment and/or would have had a different history than those of the asteroids.

The estimated porosity inside Phobos corresponds to a macroporosity content, i.e. space of voids between the blocks of rocks composing the moon (Britt et al. 2002). It is an additional porosity to the microporosity, which is on average 20% and less than 10% of the volume of the grain for carbonaceous and silicate chondritic material, respectively (Consolmagno et al. 2008), and up to 40% of the volume of the grain for the Tagish Lake meteorite as quoted above. Andert et al. (2010) have proposed to compute the porosity inside Phobos by taking into account the grain density of the material analog instead of its bulk density,[19] yielding a range of 25% to 45% of the volume occupied by voids for the range of plausible material analog composing Phobos. The Stickney crater on Phobos is supporting large macroporosity inside Phobos (Andert et al. 2010). Indeed, the formation of large craters on a small body (about the diameter of the body) requires such a large porosity in the interior in order not to destroy the body during the impact process (Richardson et al. 2002). A highly porous body is also expected to be less resistant to Mars' tidal forces than a solid body, thus preventing it to closely orbit Mars. The Roche limit of a highly porous Phobos (with 30% of macroporosity) is expected to be at a distance to Mars of about 2 R_M (Sharma 2009), therefore compatible with the current distance of Phobos to Mars (about 2.76 R_M, see Table 1).

The density of Deimos is lower than the density of Phobos, thus the macroporosity estimates are larger for the same range of material analogs (35% and up to about 60% of the volume occupied by voids for hydrated carbonaceous and dense silicate material, respectively, see Fig. 4). Note that, unlike for Phobos, the density of Deimos is lower than the density of the Tagish Lake material. Thus, a Deimos made of this material would require between 0% and 25% of macroporosity in its interior to account for its observed bulk density within its error bars (Fig. 4). However, the same contradiction as for Phobos is raised from the comparison between macroporosity estimates of Deimos and those of asteroids.

The large macroporosity content estimated inside Phobos and Deimos seems to indicate that their interiors do not correspond to the interior of a monolithic rocky body but to a gravitational aggregate of loosely consolidated material (Richardson et al. 2002) instead.

4.2 Water ice inside Phobos and Deimos

Although no evidence of hydration of the surfaces of Phobos and Deimos has been observed, it cannot be precluded that their interiors contain some amount of water ice. Indeed, the temperature conditions inside Phobos allows the presence of water ice in its interior (Fanale and Salvail 1989, 1990) and the regolith covering the surface may protect the water ice from sublimation and transport through the surface in large quantities.

[19] This porosity estimate corresponds to a macroporosity estimate assuming the material does not contain any microporosity.

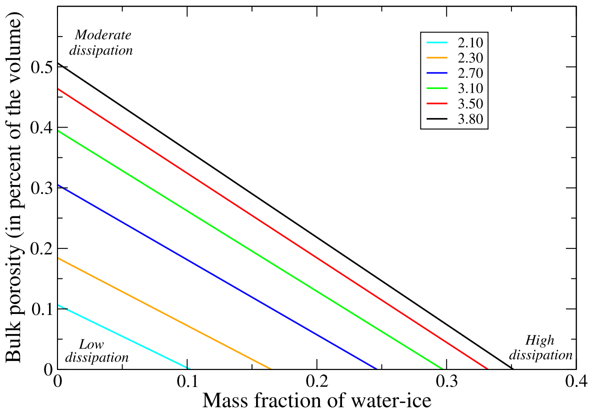

Fig. 5 Estimation of the bulk porosity and water ice content inside Phobos that fit its bulk density for a large range of rocky material grain densities (as labeled in the legend box). The expected dissipative properties of the interior of Phobos are qualitatively indicated since the quantitative details remain to be computed (see text for details)

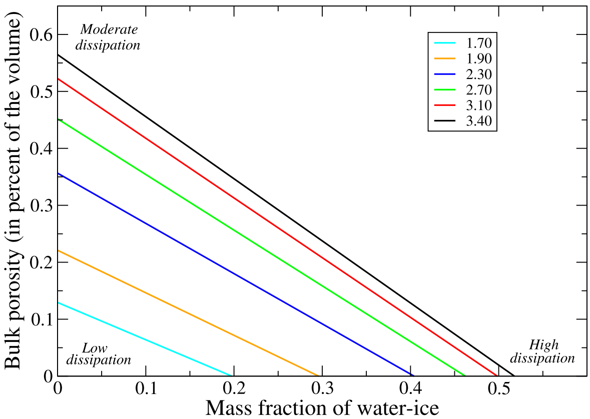

Fig. 6 Same as for Fig. 5 but for Deimos

If one assumes that both moons are composed of rocky and water ice materials, the possible amount of water ice in their interior can be computed from their density as done for the porosity content. Water ice could represent between about 10% and 35% of the mass of Phobos, depending on the density of the rocky material analog (see values on x-axis of Fig. 5). However, the upper part of this range is obtained for dense silicate material or black chondrite material. This material is expected to be condensed in the inner solar nebula (Gradie and Tedesco 1982) at a heliocentric distance of about less than 2.5 AU, and thus is expected to contain a low amount of volatiles like water ice. For rocky material of lower density, like for hydrated carbonaceous chondritic material (expected to be formed beyond about 3.5 AU), the amount of water ice is expected to be larger. Nevertheless, to fit Phobos' bulk density with such a mixture of light rock (density lower than about 2 g/cm^3) and water ice materials, no more than 10% of its mass would be composed of water ice (see values on x-axis of Fig. 5). Such a low amount of water ice seems difficult to reconcile with bodies formed near the Jupiter orbit or beyond, like T or D-type asteroids, because at such heliocentric distances, frozen volatiles like water ice, would compose about one-third of the mass of the objects (Consolmagno et al. 2008). As Deimos' density is lower than Phobos' density, the water ice content in its interior is larger for the same range of rocky material densities. For example, water ice is estimated to contribute for about 20% of Deimos' mass, when considering low density rocky material (see values on x-axis of Fig. 6).

The water ice content in Phobos and Deimos is lower also when porosity is considered. In that case, the volume fraction of bulk porosity Φ_b is related to the mass fraction of water ice f_{ice} by the following relationship:

$$\Phi_b = 1 - \frac{\rho_b}{\rho_g} - f_{\text{ice}}\rho_b \frac{(\rho_g - \rho_{\text{ice}})}{\rho_g \rho_{\text{ice}}} \tag{3}$$

where ρ_g and ρ_b are the grain density of the rocky material and the bulk density of the moon, respectively, and ρ_{ice} (0.97 g/cm^3) is the water ice density. The bulk density of the moons alone cannot permit to infer the relative porosity/water ice content in their interior, and the water ice content given above for zero porosity, is actually the maximum possible water ice content in both moons for the considered range of material densities (Figs. 5 and 6).

4.3 Porosity, water ice and origin of the Martian moons

4.3.1 Porosity and origin

Porosity and capture scenario If Phobos and Deimos are captured asteroids, their porosity may be explained by repeated impacts at their surfaces (Asphaug et al. 2002). This porosity would correspond to macroporosity in the form of large fractures inside their interior and would not exceed about 15% of the volume (Britt et al. 2002), which is consistent with the macroporosity estimates for Phobos, if it is composed of hydrated carbonaceous material (Fig. 3), and for Deimos, if it is composed of the Tagish Lake material (Fig. 4). It has been proposed that the grooves on Phobos might be surface expressions of deep-seated fractures in its interior, however, numerous other theories have been proposed to account for the grooves, which do not necessarily require such an internal structure (see for example recent works by Murray et al. 2006; Hamelin 2011).

A larger porosity would require that a complete shattering and reassembling event occurred in the history of the moons, as thought to have occurred in the history of many asteroids due to violent collisions in the asteroid belt (Britt et al. 2002; Richardson et al. 2002). Indeed, after the collision, the larger debris reassemble first due to their larger gravitational attraction, forming a core of large boulders with large gaps between them. Then, the smaller debris reassemble, but do not fill the gaps because of the low self-gravity of the reassembling body. The large gaps are left in the volume of the reassembled body, explaining a high porosity (macroporosity) content in its interior (Richardson et al. 2002).

If Phobos and Deimos are porous asteroids captured by Mars, it still requires that their interior could dissipate enough orbital energy by tidal effects to account for their current orbits around Mars. Therefore, the question arises whether a porous body can have a significantly larger tidal dissipation rate than a pure rocky body. Recently, Goldreich and Sari (2009) have estimated the rigidity of highly macroporous (30% of porosity) small-sized bodies, and found it could be lowered by a factor of up to 10 with respect to a pure rocky body. In turn, the k_2 Love number of the macroporous body is increased by the same factor (Goldreich and Sari 2009). Applied to the case of a macroporous Phobos, it means that the rate of orbital changes rate increase by the same factor of 10, for a fixed Q value. On another hand, the increase of k_2 would provide the same orbital changes rate with a Q value increased by the same factor of 10. In order to get significant orbital changes as in Mignard (1981), the Q value of such macroporous Phobos would be about 20 (see Sect. 2.2.3). However, this value seems to be still too low for a macroporous rocky object. Indeed, experimental measurements on granular material extrapolated to Phobos' conditions seem to indicate that the Q value is still larger than 100 (Castillo-Rogez et al. 2011).

The microporosity can also lower the rigidity of rocky material, but more than 50% of microporosity is required to lower it by a factor of 10 (Jaeger et al. 2007). The material analogs proposed for Phobos, so far, do not contain such a high content of microporosity.

The porosity is thus expected to slightly enhance the dissipative properties inside Phobos and Deimos, thus increasing the rate of the tidally induced orbital changes, although not so much as required by the capture scenario.

Porosity and in situ formation The scenario proposing that Phobos and Deimos are remnants of a former larger moon captured and then destroyed by Mars' tidal forces is not consistent with a high porosity inside the moons. Indeed, in this scenario, the former moon is large enough to significantly remove porosity in its interior. In turn, the remnants are not expected to have large porosity in their interior. However, this scenario can be reconciled with a high porosity content inside Phobos and Deimos, if these two moons were formed by reassembling of smaller remnants (the high porosity coming from the reassembling process as for the reassembly of asteroid debris after violent collisions, Richardson et al. 2002). However, this scenario has not been studied in detail yet, to confirm whether it is plausible or not.

The high porosity is consistent with the scenario of moon formation in a Mars' circum accretion disk resulting from the collision between Mars and a massive former body (Craddock 2011). Indeed, recent theoretical works about accretion processes have shown that small-sized porous bodies can be formed around Saturn from gravity instabilities in the rings of the planet, providing a new scenario of origin for some Saturn's small moons (Charnoz et al. 2010). The high porosity inside the Martian moons thus indicates a new way of research by applying the driving mechanisms of accretion disk to the case of Phobos and Deimos (Rosenblatt and Charnoz 2011). High porosity inside Phobos and Deimos provides a new support to the scenario of formation of the Martian moons from an accretion disk around Mars. In addition, as argued by Giuranna et al. (2011), the possible silicate composition of Phobos also supports that scenario. Note, however, that a carbonaceous composition of Phobos and Deimos is not excluded in this scenario, since the composition of the debris blasted into Mars' orbit by the collision may come from the impactor, which could be of carbonaceous composition.[20]

[20] In the early solar system, volatile-rich 'embryos' objects could have migrated from the outer to the inner solar system (Lunine 2006). The impactor involved in the scenario proposed by Craddock (2011) might have been such a planetary embryo.

4.3.2 *Water ice and origin*

Water ice is a possible component in the interior of Phobos and Deimos, which is pertinent for their origin. Indeed, water ice can increase the dissipation rate inside a rocky body, by at least two orders of magnitude, especially if melted ice is filling the pore spaces in the rocky matrix (McCarthy and Castillo-Rogez 2011). This perspective is especially interesting in the case of Deimos, for which the tidally induced orbital changes have been shown to insufficiently change its orbit over the last 4.6 Ga (e.g. Burns 1992). Further investigations are, however, needed in order to assess how much the possible water ice inside the moons could increase the dissipative properties of their interiors, and in turn, the tidally induced orbital changes in order to assess whether the orbital changes required by the capture scenario might be satisfied or not (see Sect. 2.2.3).

In the in situ formation scenario proposed by Craddock (2011), one may expect a low water ice content inside the accreted bodies in Mars' orbit. Indeed, in this scenario, the debris blasted in Mars' orbit would have been completely melted by the heat released from the collision. During the cooling of the disk, the volatiles compounds of the debris may not recondense with the rocky compounds, depending on the thermodynamic conditions in the disk. That point deserves further investigations as has been done in the case of the formation of the Moon of the Earth (e.g. Canup 2004).

5 Perspectives for future missions: the internal structure of the moons as a key observational constraint for their origin

In spite of the new data collected by the Mars Express mission, the question of the origin of the Martian moons remains unanswered. However, some of these data emphasize the importance of the link between the internal structure and the origin of the moons, opening new paths of investigation. In particular, the Mars Express data have permitted to precisely determine the density of Phobos (yielding a new view of the interior of this small body). On the one hand, Phobos' interior can sustain a high porosity, which raises a renewed interest for the in situ formation scenarios (Andert et al. 2010; Rosenblatt et al. 2010). On the other hand, the density can also be explained by a water-rich interior of Phobos with important consequences concerning the capture scenario. Indeed, water ice is expected to significantly increase the tidal rate dissipation inside Phobos, and in turn, to modify the previous results about the past evolution of Phobos' orbit by tidal effects. However, this latter point needs further investigations in order to first assess a realistic tidal dissipation rate inside Phobos (made of rock and water ice, Castillo-Rogez et al. 2011) and then to explore the consequences in terms of tidally induced orbital evolution. The density alone cannot provide tight constraints for this purpose since the water ice content also depends on the porosity inside Phobos (Fig. 5). More data about the interior are thus needed in order to constrain the respective proportion of porosity and water ice and how these light compounds might be distributed in the volume of this small body (Rosenblatt et al. 2010).

The Mars Express images of Phobos have also been used to determine the amplitude of the forced libration in longitude (or periodic variations of its spin rate) as $1.24° +/- 0.15°$ (Willner et al. 2010). Since the error bar includes the expected value of 1.1 ° obtained from the shape of Phobos in the case of homogeneous mass distribution, this measurement cannot distinguish between a homogeneous and heterogeneous mass distributions in Phobos. The error bar indicates, however, a slightly heterogeneous Phobos interior, as suggested by recent models of internal mass distribution (Rosenblatt et al. 2010). The same models also show that the values of the principal moments of inertia of Phobos slightly vary, depending on the content and distribution of porosity and water ice in the volume of Phobos. This implies that the precise measurement of these moments of inertia may provide tighter constraints on the interior structure of Phobos (Rosenblatt et al. 2011). As the moments of inertia are related to the forced libration amplitude and to the second-order coefficients of the non-spherical part of the gravity field of Phobos (Borderies and Yoder 1990), the precise measurement of this gravity field and libration amplitude would provide precise measurement of the moments of inertia of Phobos. However, the measurement of the gravity field is challenging (Andert et al. 2011) since it needs very close flybys (with a closest approach of about 50 km). The extended phase of the Mars Express mission (up to 2014) would offer opportunities of such very close flybys.

The Russian mission Phobos–Soil is due for launch in November 2011, and arrival at Mars in August 2012 (e.g. Zelenyi and Zakharov 2011; Martynov and Khartov 2011). The primary goal of this mission is to bring samples of the soil of Phobos back to Earth. These samples will permit to determine the ‘true’ composition of the surface of Phobos. It will also permit to better characterize the space weathering effect in the Martian environment as it is needed for better interpreting the remote-sensing data in terms of the composition of Phobos (and Deimos) surface. The Phobos soil samples will also allow for analyses of radio-isotopic ratios such as those of oxygen. These geochemical analyses would permit, by comparison with those of SNC[21] meteorites thought to come from Mars’ surface, to know whether Phobos and Mars stem from the same geochemical reservoir or not. The same analyses performed on the Lunar samples of the Apollo missions have revealed the strong similarity between the Earth and the Moon, which led to the current explanation of the origin of the Moon (Hartmann 1976).

The Phobos–Soil mission will only sample the soil of Phobos, which might not be representative of the bulk composition of Phobos, and so might not provide direct tighter constraints on its internal structure. Investigations on the interior of Phobos will thus still be needed in order to support the interpretation of the sample analyses. In the first phases of its mission (in January 2013), the Phobos–Soil spacecraft will orbit Mars at very close distances to Phobos (45–55 km) before it lands on Phobos in February 2013. During the orbits close to Phobos, the spacecraft will be able to sense the gravity field of Phobos, especially the second-order coefficients (Rosenblatt et al. 2010). Once landed on Phobos’ surface, the spacecraft will also be used to measure the fine variations of its rotation and of the orientation of its axis of rotation. Both gravity field and rotation variation measurements will provide for the first time

[21] Shergottites, Nakhlites, Chassignites.

the opportunity to precisely measure the moments of inertia of Phobos, and thus to provide a better understanding of its internal structure (e.g. Castillo-Rogez et al. 2011; Le Maistre et al. 2011; Rambaux et al. 2011; Rosenblatt et al. 2011).

Both ongoing Mars Express and soon-coming Phobos–Soil missions will provide a unique dataset of the surface and interior of Phobos. Projects of future missions to the Martian moons (Oberst et al. 2011; Michel et al. 2011) could also improve such a dataset, especially for Deimos. The comparison of Martian moon material samples with asteroid material samples, gathered by future missions to asteroids such as Marco Polo, could also be helpful. The results of these current and future missions will lead us, perhaps, to answer the question of the origin of the Martian moons.

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