The Syrtis Major volcanic province, Mars: Synthesis from Mars Global Surveyor data

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[1] We investigated the geology, stratigraphy, morphology, and topography of Syrtis Major, one of the large Hesperian-aged volcanic provinces on Mars. New data from the Mars Global Surveyor (MGS) and Odyssey spacecraft allowed us to study Syrtis Major in unprecedented detail. On the basis of Mars Orbiter Laser Altimeter (MOLA) data we estimated the thickness of Syrtis Major lavas to be on the order of \(0.5-1.0\) km, their volume being \(1.6-3.2 \times 10^5\) km\(^3\). MOLA data also show that the calderas of Nili and Meroe Patera are located within a large N-S elongated central depression and are on the order of 2 km deep. MOLA data further indicate that the slopes of the flanks of the shield are \(\ll1^\circ\). These very gentle slopes are similar to other Martian highland paterae such as Amphitrites, Tyrhena, and Syria Planum but are much smaller than slopes measured on the flanks of large Martian shield volcanoes such as Olympus, Elysium, Asraeus, and Arslia Mons. At kilometer-scale baselines, the surface of the Syrtis Major Formation is smoother than the surrounding highland plains but rougher than the floor of Isidis. We propose that the differences in surface roughness and wrinkle ridge patterns between Isidis Planitia and Syrtis Major are probably caused by additional sedimentation in the impact basin. Both the wrinkle ridge pattern and topography of Syrtis Major are interpreted to be consistent with a gravitational collapse of the southeastern sector of Syrtis Major. Recently a model has been proposed to explain the low elevation of the Isidis rim in the Syrtis Major region [e.g., Tanaka et al., 2002]. This model calls on catastrophic erosion of the rim that is triggered by the emplacement of sills, which interacted with a volatile-rich substrate. We used MOLA data from nearby terrains of the Isidis rim to investigate the Isidis rim in locations where it has supposedly been eroded. On the basis of our investigation we conclude that the smooth morphology and relatively low topography of the Isidis rim in the Syrtis Major region can be best explained by initial heterogeneities in topography and flooding with lavas, together with some erosion and/or tectonic modification of the original rim. New crater counts show that Syrtis Major is of Early Hesperian age (N(5) = 154), older than in the geologic map of Greeley and Guest [1987]. On the basis of inspection of all available Mars Orbiter Camera (MOC) images we identified several terrain types and developed a stratigraphic sequence for Syrtis Major. Large-scale erosional modification (e.g., deflation) of the Syrtis Major lavas and/or tephra deposits and redistribution of eolian material are evident in numerous MOC images.


1. Introduction

[2] Syrtis Major is one of the most prominent Hesperian volcanic complexes on Mars. Simpson et al. [1982] concluded that albedo, radar properties, and low crater densities of Syrtis Major resemble those of lunar maria. Several authors [Meyer and Grolier, 1977; Greeley and Spudis, 1978] interpreted the circular shape of Syrtis Major as impact-related, but Schaber [1982] showed that the topography of the structure is more consistent with a low shield volcano. Initial analysis of MOLA data supported the interpretation of central Syrtis Major as a shield volcano [Head et al., 1998]. However, the low relief of Syrtis Major
is strikingly different from other Martian shield volcanoes (e.g., Tharsis Montes) and may be related to changes in composition, differentiation history, eruptive styles or differences in crustal thickness [Schaber, 1982]. Individual lava flows of Syrtis Major are \(~25–30\) m thick [Head et al., 1998] and are up to \(120–150\) km long [Schaber, 1982]. Data of the Phobos mission Imaging Spectrometer for Mars (ISM) suggest a composition of these lava flows that is similar to the Martian SNC meteorites (Shergottites, Nakhlites, Chassignites) [Mustard et al., 1997] and the absence of ultramafic or komatiitic lavas [Mustard et al., 2001].

3. Results

3.1. General Topography of Syrtis Major

Figure 2a shows the MOLA gridded topography of Syrtis Major and the western Isidis basin. The superposed black line outlines the extent of Syrtis Major volcanic flows (unit Hs) in the geologic map of Greeley and Guest [1987]. On the basis of the MOLA topography we see that the highest point of Syrtis Major is located NW of Nili Patera at \(~2300\) m. The calderas of Nili and Meroe Patera are located within a larger NW-SE elongated central depression and are \(~2000\) m deep. The terrain southwest, west and north of the calderas appears to be higher (\(2000–2300\) m) than the terrain east and southeast of the calderas (\(1500–1700\) m) and forms a crescent-like summit. On the basis of the extent of unit Hs in the geologic map of Greeley and Guest [1987] and the new MOLA data we can trace Syrtis Major lavas in the western Isidis basin down to at least \(~3950\) m. On the basis of analyses of Head et al. [2001], Ivanov and Head [2002] and our own study [e.g., Hiesinger and Head, 2002], it appears that Syrtis lavas have covered parts of the Isidis basin floor and underlie units Hv1 and Hv2. Ivanov and Head [2002] also identified young lava flows, which are superposed on the Isidis units, suggesting a complex interaction of Syrtis lavas with the Isidis units. The northern boundary between Syrtis Major and surrounding highlands is significantly lower in elevation (\(~300–700\) m) than the boundary along the southern margin of Syrtis Major (\(~800\) to \(~1800\) m) and we observe a very level area at \(~1600\) m in the southwestern parts of Syrtis Major (Figure 2a).

3.2. Slopes

A slope map based on MOLA data indicates significant differences between the Noachian highlands, the Hesperian Syrtis Major lavas, and the units associated with the Isidis basin (Figure 3a). This map was created on the basis of 32 pixel per degree MOLA data and shows slopes at a baseline of \(~5.6\) km. The geologic maps of this area [Meyer and Grolier, 1977; Greeley and Guest, 1987], together with MOLA data, indicate that Syrtis Major lavas are superposed on northward tilted Noachian plains units. Slopes of the Syrtis lavas are less steep than of the Noachian highlands, but are significantly steeper than of the Isidis units. Steepest slopes of Syrtis Major are associated with crater rims, the central depression, the calderas, and the transition into the Isidis basin. These slopes are on the order of \(1–5\)° but can locally be as much as \(10\)° (Figure 3a). We observe steeper slopes into the summit depression in the west (\(3–5\)°) compared to the east (\(1–2\)°). Large areas of Syrtis Major have regional slopes of much <1°. Our slope map also indicates...
that western parts of Syrtis Major and crater Antoniadi exhibit less steep slopes (<0.5 degree) than the rest of Syrtis Major, especially than the passage into the Isidis basin (>1 degree). Slopes into the Isidis basin are less steep in the north than in the south and there is a distinctive cliff of ~500 m height in the north.

[8] Detailed investigation of MOLA data shows that regional slopes of the Syrtis Major lavas to the north are ~0.13°, to the south are ~0.02°, and slopes to the west are on the order of 0.08° over baselines of hundreds of kilometers. We observe slopes of ~0.4° to the east, that is into the Isidis basin (Figure 4). These slopes reflect the low profile of Syrtis Major and are significantly smaller than for other Martian volcanoes, such as Olympus Mons, Alba Patera, and Elysium Mons [Head et al., 1998] (Figure 5). However, as Figure 5 shows, Syrtis Major slopes are similar to highland paterae such as Amphitrites or Thyrenna Patera and Syria Planum.

3.3. Surface Roughness

[9] On the basis of a MOLA map of the kilometer-scale surface roughness (Figure 3b), Syrtis Major can be clearly distinguished from the surrounding cratered highlands that are generally rougher at all wavelengths than the Syrtis Major Formation (Hs). The roughness map shows that the Syrtis Major Formation is rougher at all wavelengths (0.6–19.2 km) compared to other investigated Martian volcanic units (e.g., Hr, At₄, At₅, Aop) [Kreslavsky and Head, 2000]. We found that western and northwestern parts of Syrtis Major are probably slightly smoother at all wavelengths compared to eastern and southeastern areas as indicated by darker hues in the surface roughness map of Kreslavsky...
and Head [2000] (Figure 3b). This difference in surface roughness is largest for short wavelengths and less prominent for intermediate and long wavelengths, resulting in more bluish hues in the east compared to more reddish hues in the west. Kreslavsky and Head [2000] found that the median differential slope of unit Hs varies almost linearly between ~0.1° at 19.2 km wavelength and ~0.2° at 0.6 km wavelength. This linear variation in median differential surface roughness is a characteristic of volcanic plains units and is different from the units of the Vastitas Borealis Formation (HvK, Hvm, Hvr, Hvg), which are all characterized by a dominance of intermediate-scale roughnesses. One member of the Vastitas Borealis Formation, unit Hvr, is exposed on the floor of the Isidis basin and is characterized by median differential slopes of ~0.04° at 19.2 km baseline length, ~0.12° at 2.4 km baseline, and ~0.1° at 0.6 km baseline [Kreslavsky and Head, 2000]. At short wavelengths (0.6 km) Isidis Planitia appears rougher than Syrtis Major; at long wavelengths (19.2 km) Isidis Planitia is smoother than Syrtis Major and this might be related to sedimentation within the impact basin [e.g., Greeley and Guest, 1987; Grizzetti and Schulz, 1989; Head et al., 2001; Head and Bridges, 2001; Bridges et al., 2003].

Aharonson et al. [2001] used MOLA data to globally calculate the RMS (root mean squares) and median slopes in 35-km windows. On the basis of their plate 2, we see a well pronounced difference in median slope between eastern and western parts of Syrtis Major. According to their map western parts of Syrtis Major are characterized by median slopes of ~0.25–0.45°; eastern parts are rougher and exhibit slopes of ~0.45–0.7°, which are similar to or slightly smaller than slopes of the floor of the Isidis basin (~0.5–0.75°).

Smith et al. [2001] report that MOLA data have been used to calculate surface slopes on baselines from 300 m to 1000 km [e.g., Kreslavsky and Head, 2000; Aharonson et al., 2001]. At even shorter baselines, MOLA optical pulse widths can be used to measure the surface roughness at 100-m-scale baselines [e.g., Garvin et al., 1999; Smith et al., 2001]. This is possible because MOLA measures the effective width of the backscattered laser energy. The broadening of the energy beam due to interactions with the Martian surface can be used to calculate RMS vertical surface roughnesses [Garvin et al., 1999]. On the basis of MOLA optical pulse width data, the mean 100-m-scale global roughness is 2.1 ± 2.0 m RMS [Garvin et al., 1999]. The roughness data show a bimodal distribution with the primary mode at 1.5 m RMS and <1% of the values in excess of 10 m RMS. In the pulse width surface roughness map of Smith et al. [2001], Syrtis Major exhibits a vertical roughness of <1.5 m RMS and we do not see differences in small-scale surface roughness between east and west. In the same map, the floor of Isidis appears to be rougher with vertical roughnesses of ~2 m RMS.

3.4. Gravity and Crustal Thickness

On the basis of MOLA data, Smith et al. [1999b] and Zuber et al. [2000] derived maps of the global gravity field of Mars. These maps show large positive gravity anomalies associated with some of the major impact structures such as Isidis, Utopia, or Argyre, and the Tharsis volcanic complex [Smith et al., 1999b; Zuber et al., 2000]. Smith et al.
Figure 4. Shaded-relief map of Syrtis Major based on MOLA gridded topography data with superposed (a) East-West profiles and (b) North-South profiles. Vertical exaggeration of the profiles is ~40x.
also reported positive gravity anomalies associated with moderate-diameter (100 km) impact basins in the northern lowlands. Syrtis Major, 1100 km in diameter, is practically indistinguishable from the surrounding highland terrain and there is no evidence for a major gravity anomaly associated with Syrtis Major. Schaber [1982] argued that the absence of a gravity anomaly is inconsistent with the existence of an impact basin prior to the onset of volcanism in the Syrtis Major region. The gravitational signature of Syrtis Major is similar to other highland volcanic complexes such as Amphitrites, Tyrrhena, and Hadriaca Patera for which no impact-related origin has been proposed. However, theoretically it remains possible that the potential basin might have been compensated to just the degree where neither a positive nor a negative gravity anomaly can be resolved with Mars Global Surveyor (MGS) data. This appears unlikely because the nearby Noachian Isidis basin shows a prominent positive gravity anomaly of 600 mgal [Smith et al., 1999b]. From this discussion we conclude that, based on gravity data alone, the hypothesis of an impact occurring prior to the emplacement of Syrtis Major can neither be unambiguously proven nor rejected. However, as all large Martian impact basins (e.g., Isidis, Argyre, Hellas, Utopia) and some moderate-diameter (100 km) impact basins in the northern lowlands show gravity anomalies it seems unlikely that the origin of Syrtis Major is impact-related.

A map derived from MGS gravity and topography data does not show significant differences in crustal thickness underneath the Syrtis Major volcanic complex compared to adjacent cratered highland plains [Smith et al., 1999b; Zuber et al., 2000]. Cratered highlands and Syrtis Major both have a crustal thickness on the order of ~45–60 km and Syrtis Major is practically indistinguishable from the surrounding highland terrain. However, the crust is significantly thinner compared to the Tharsis area (~60 km). This suggests that Syrtis Major lavas are rather thin compared to other Martian volcanic complexes such as Tharsis and Alba Patera. This interpretation is consistent with independent thickness estimates based on MOLA topography and the absence of a positive gravity anomaly associated with Syrtis Major, which suggests that the mass contribution of the Syrtis Major lavas is trivial compared with the contribution of the pre-volcanic crust to the gravity signature.

### 3.5. Thickness and Volume of Syrtis Major Lavas

According to Schaber [1982] and Hodges and Moore [1994] lava deposits of Syrtis Major are ~1100 km in diameter and have an estimated maximum thickness of only 0.5 km. On the basis of the assumption that the mean ridge spacing reflects the dominant wavelength of folding, Saunders et al. [1981] estimated the thickness of the ridged plains in Syrtis Major to be ~0.9 km. E-W profiles based on MOLA data show that the height of the shield is ~0.5 km (Figure 6), such being consistent with previous estimates of Schaber [1982] that were based on radar data. MOLA N-S profiles indicate a slightly higher edifice just under 1 km in height (Figure 6). For our estimates of the thickness of the Syrtis Major lavas we correlated MOLA data with the lateral extent of unit Hs as mapped by Greeley and Guest [1987]. We then interpolated linearly between the elevations of the most extreme extents of unit Hs as mapped by Greeley and Guest [1987]. We then interpolated linearly between the elevations of the most extreme extents of unit Hs, both in N-S and E-W directions (Figure 6). This baseline was used to calculate the difference to the highest point of Syrtis Major, which gives us the minimum thickness of the Syrtis Major lavas. Independent thickness estimates that we performed by making use of the relationship between diameter and rim height of buried impact craters (still visible as circular wrinkle ridges), indicate that in areas with circular wrinkle ridges the lavas are <600 m thick (section 3.7). In areas without circular wrinkle ridges the thickness of the lavas might be greater. Wichmann and Schultz [1988] argued that in order to bury the massif ring of the Isidis basin a thickness of 1–2 km with a maximum thickness of 2.5 km is required. This implies that the Isidis rim did still exist at times of the lava emplacement. However, the
Isidis rim might have never existed in the form envisioned by Wichmann and Schultz [1988] or might have been eroded prior to the emplacement of the Syrtis Major lavas as recently suggested by Tanaka et al. [2001a, 2002]. In section 3.10 we will discuss the relationship between Syrtis Major lavas and the rim of the Isidis basin in greater detail.

[15] On the basis of our estimates of the thickness of the lavas of 0.5–1.0 km (Figure 6) and a diameter of 1100 km, we calculated the volume of Syrtis Major lavas to be $1.6–3.2 \times 10^5$ km$^3$, assuming a very low, cone-like morphology of the volcanic shield. These volumes are comparable to estimates of Schaber [1982], which are on the order of $2 \times 10^5$ km$^3$. For comparison we provide a very rough estimate of the volume of Amphitrites Patera, which is based on a diameter of $\sim 1000$ km and a height of $\sim 1.2$ km. On the basis of these dimensions, the volume of Amphitrites Patera would be on the order of $3.2 \times 10^5$ km$^3$, hence very similar to the volume of Syrtis Major. However, because the extent of Amphitrites Patera is only poorly defined [Hodges and Moore, 1994] and its location on the southern rim of the Hellas basin, the volume of Amphitrites Patera is difficult to estimate and might be subject to significant errors.

3.6. Composition of Syrtis Major Lava

[16] In the past extensive research has been done on the composition and mineralogy of Syrtis Major lavas [e.g.,...
been pointed out by several authors [e.g., Erard et al., 1990; Mustard et al., 1993, 1997; Reyes and Christensen, 1994; Bandfield et al., 2000; Noble and Pieters, 2001; Wyatt et al., 2002; Wyatt and McSween, 2002]. Unfortunately, the interpretation of the available data led to a wide variety of possible compositions, ranging from basalts to andesites to komatiites.

[17] Mustard et al. [1993] used Phobos-ISM data in order to derive the composition of the Syrtis Major lavas. They determined that the mafic mineralogy is dominated by augite-bearing pyroxenes, and minor amounts of olivine, feldspar, and glass. From their data they concluded that ultramafic or komatiitic lavas are absent on the surface of Syrtis Major [Mustard et al., 1993]. However, Reyes and Christensen [1994] argued that ISM data are consistent with a komatiitic composition of the Syrtis lavas. Mustard et al. [1997] concluded that Syrtis Major lavas are dominated by a two pyroxene mineralogy (low and high calcium), that the Martian mantle was depleted in aluminum at times of melt production, and that there was little evolution in the composition of the mantle source regions between >3 b.y. and 180 m.y. On the basis of their investigation of imaging spectrometer (ISM) data, Mustard et al. [1990] also reported differences in spectral characteristics between an eastern and a western unit of Syrtis Major; the eastern unit having a decreasing reflectance toward longer wavelengths, the western unit having a flat spectrum. They interpreted the majority of spectral variation across Syrtis Major to be related to mixing of volcanic bedrock and debris with highly altered dust and soil [Mustard et al., 1993]. This mixing ranges from aerial mixing in the western parts to nonlinear intimate mixing or convolution of spectral properties through thin coatings in the eastern parts [Mustard et al., 1993]. Differences in the distribution of large-scale wind streaks between the east and west, as we will describe in section 3.12, also might be related to such variations in mixing characteristics. On the basis of spectroscopic investigation of analog materials, Harloff and Arnold [2002] reported that bulk samples show a preference for a negative near infrared (NIR) continuum slope, in contrast to powder samples. They concluded that this suggests a different interpretation of ISM data, namely that the very dark regions, which are characterized by a flat NIR continuum, are covered by sand-sized basaltic particles. Dark regions, which are characterized by a blue continuum slope are dominated by bedrock or blocky basaltic rocks [Harloff and Arnold, 2002].

[18] On the basis of TES data, Bandfield et al. [2000] produced global maps of the distribution of two spectral end-member types, a basaltic (Type 1 terrain) and an andesitic type (Type 2 terrain). We superposed their maps of the andesite and basalt concentration on the MOLA topography (Figure 7a). The unit they interpreted as andesite preferentially occurs along the western and northern edge of Syrtis Major. Highest concentrations of the unit they interpret to be basalt are mainly found in central parts of Syrtis Major, north and south of the calderas. However, it has been pointed out by several authors [e.g., Noble and Pieters, 2001; Wyatt et al., 2002; Wyatt and McSween, 2002] that the identification of andesite on the Martian surface is subject to alternative interpretations. Noble and Pieters [2001] concluded that the TES spectra of Type 2 terrain are equally consistent with (1) an interpretation as andesite as originally proposed by Bandfield et al. [2000], (2) a basalt with a larger glass component that may or may not differ compositionally from Type 1 terrain, and (3) weathering processes that have altered the surface to resemble the spectral characteristics of andesite. In this case the composition of Type 1 and Type 2 terrains may or may not differ [Noble and Pieters, 2001]. Wyatt and McSween [2002] found that some clays and high-silica glass, which has been used in the original analysis, are spectrally similar. They concluded that Type 2 terrain can be interpreted as weathered basalt, hence supporting the conclusions of Noble and Pieters [2001]. Wyatt and McSween [2002] suggested that Type 2 terrain are basalts that were weathered under submarine conditions or are sediments derived from weathered basalts and deposited in the northern lowlands. Wyatt et al. [2002] found a gradational boundary between Type 1 and Type 2 terrains in Acidalia and Chryse Planitiae and concluded that it may represent (1) an influx of basaltic sediments that originated from the southern highlands and were deposited on andesitic volcanics, or (2) incompletely weathered basalts that mark the geographic extent of submarine alteration of basalts. However, submarine alteration of basalts is unlikely for Syrtis Major because this region is at least ~3 km above all proposed shorelines [e.g., Parker et al., 1989; Parker et al., 1993; Edgett and Parker, 1997].

[19] Bandfield et al. [2000] argued that basalts are restricted to the more distant past and that andesites occur throughout Martian history and more dominantly in more recent times. On the basis of their analysis of small-scale isolated basalt occurrences in the northern lowlands (Cerberus, Milankovich crater, Pettit crater, Erebus Montes), Rogers and Christensen [2002] suggested that all of the Syrtis Major-type basalt may be older than the andesite. If Bandfield et al. [2000] and Rogers and Christensen [2002] are correct, then the andesites at the western and northern edges of Syrtis Major could represent some late stage flank eruptions. Alternatively, Type 2 material could exhibit a larger glass component or simply be more heavily weathered basalt [e.g., Noble and Pieters, 2001; Wyatt et al., 2002; Wyatt and McSween, 2002].

[20] While the gamma-ray spectrometer (GRS) on board Mars Odyssey found ample evidence for water at high latitudes, there is no evidence for similarly large amounts of hydrogen in the upper 1 m of the soil of Syrtis Major or the Isidis basin [Boynton et al., 2002]. Interestingly, epithermal neutron counts are slightly higher for the Isidis basin, indicating less water, compared to Syrtis Major. Highlands west of Syrtis Major show lower epithermal neutron flux rates and therefore must contain more hydrogen/water than Syrtis Major and Isidis. These findings are confirmed by the High Energy Neutron Detector (HEND) of Mars Odyssey, which measures the fast neutron flux [Mitrofanov et al., 2002]. Erard et al. [1990] measured the depth of the 2.9 μm H₂O absorption band of ISM spectra and found Syrtis Major lavas to be very dry (25% absorption) compared to very hydrated Isidis materials (40% absorption). This observation is inconsistent with results from the gamma-ray and high energy neutron detector, which indicate that the upper 1 m of the soil is drier in Isidis than in the Syrtis Major region. Erard et al. [1990] also reported that eastern parts of Syrtis Major are drier than the western part. This finding could be related to drainage of
the eastern portions of Syrtis Major into the Isidis basin and larger amounts of pore water trapped in sediments of a formerly existing lake within the Isidis basin compared to the volcanic material that comprises Syrtis Major. On the other hand, inspection of Viking MDIM data reveals numerous craters with lobate ejecta blankets that postdate the emplacement of unit Hs. Such lobate crater shapes have been cited as evidence for a volatile-rich substrate [e.g., Carr et al., 1977; Kuzmin et al., 1988]. The map of Kuzmin et al. [1988] indicates that the onset diameter of lobate impact craters on Syrtis Major and in the Isidis basin is 4–6 km. On the basis of inspection of Viking MDIM data, we found onset diameters of ~7 and ~8 km, respectively. Interestingly, we observed smaller onset diameters and a larger number of lobate craters for Syrtis Major compared to Isidis Planitia. Squyres et al. [1992] argued that it is difficult to interpret the onset diameters in terms of absolute depth to the top of the volatile-rich layer and its volatile content. They proposed that the thickness of the desiccated upper surface layer is ~300–400 m at the equator and 200–250 m at 30° latitude. On the basis of the experimental laboratory work, Stewart et al. [2001] reported that only 5–15% vol water/ice are sufficient to produce lobate crater morphologies that closely match morphologies observed with MOLA. If it is true that only small amounts of water are needed to produce lobate ejecta morphologies, it is not clear why the Isidis basin, which is a potential regional sink for volatiles, shows fewer lobate craters than Syrtis Major. One option is that lobate ejecta blankets are formed by atmospheric effects and independently of the volatile content of the substrate [Schultz, 1986].

[21] From this discussion we conclude that Syrtis Major has to be further investigated, for example with high-resolution THEMIS data, in order to place better constraints on its composition and mineralogy.

3.7. Wrinkle Ridges

[22] A shaded-relief map, based on grided MOLA topography, exhibits numerous radial, concentric, and circular features that we interpret as wrinkle ridges, flow fronts, and lava channels. However, arcuate ridges in eastern parts of Syrtis Major have also been interpreted as remnants of a large mudflow that emerged from the eastern flanks of the Nili/Meroe Patera shield [Jöns, 1987]. We investigated the cross-sections of ~15 wrinkle ridges on Syrtis Major in detail and found that these ridges are on average ~32 km wide (~10–60 km) and ~162 m high (50–400 m). Measurements of 6 wrinkle ridges in the adjacent Isidis basin indicate that these ridges are wider (~67 km) but less high (~122 m). These data are consistent with the hypothesis that Late Hesperian and Amazonian sediments (Hvr, Aps) in the Isidis basin, covered underlying Early Hesperian (Hr) wrinkle ridges [e.g., Head et al., 2002]. Such sedimentation would occur preferentially in the topographic lows between the ridges, would decrease the height of the wrinkle ridges, and would also increase the width and spacing of these ridges as measured with MOLA (Figure 8). Compared to the isotopic pattern of wrinkle ridges on the floor of Isidis [Head et al., 2001; Head and Bridges, 2001], we see that for major parts of Syrtis Major the scale of the pattern formed by ridges is smaller. On Syrtis Major we measured the diameter of ~50 “cells” that are outlined by ridges in order to compare them to ~35 measured diameters of similar “cells” on the floor of the Isidis basin. We found that the “cells” on Syrtis Major, which are formed by semi-circular wrinkle ridges, are usually ~50 km in diameter with very few larger than ~100 km, whereas the “cells” on the floor of the Isidis basin are on average ~80 km with a few up to ~180 km in diameter. These numbers are in good agreement with the spacing of ridges in the North Polar basin, Lunae Planum and eastern Solis Planum as measured by Head et al. [2002]. Large parts of Syrtis Major exhibit shorter ridges than in the Isidis basin and locally exhibit preferred orientations. On the basis of a MOLA shaded-relief map we mapped Syrtis Major wrinkle ridges in detail (Figure 7b, Figure 9). We found characteristic differences in the occurrence and orientation of these wrinkle ridges. In the western and northeastern quadrangles of Syrtis Major wrinkle ridges are more linear and oriented radially to the central calderas (Figure 9a), in the southeastern quadrant we often found wrinkle ridges to be more arcuate and oriented concentrically to the calderas (Figure 9b). Wrinkle ridges in the east, west and south that are closer to the central depression appear to be arcuate and concentrically oriented; wrinkle ridges further away from the central depression are linear and radially oriented. Comparing the wrinkle ridge pattern of Syrtis Major with wrinkle ridge patterns on the Moon that are related to flexural response of the lunar lithosphere to mare basalt loads [Solomon and...
we see that lunar wrinkle ridges are less arcuate, are generally larger in their dimensions, and do not form relatively small-scale circular patterns like the wrinkle ridges observed on Syrtis Major. In addition, lunar impact basins with wrinkle ridges on their basalt fills are commonly surrounded by concentric linear rilles. These linear rilles are formed by extensional stress when the basalt fill caused subsidence of the basin center. The absence of linear rilles in the investigated area and the differences in wrinkle ridge morphology suggest a different mode of formation of the wrinkle ridges on Syrtis Major.

The summit of Syrtis Major is characterized by a crescent-like high topography in the western and northwestern parts and a lower topography in the southeastern part (Figure 2). Relating the topography with the occurrence of wrinkle ridges, we see that arcuate wrinkle ridges occur preferentially in the section with lower topography. One possible interpretation is that this is related to a gravitational sector collapse of Syrtis Major. In this scenario, southeastern parts of the summit collapsed due to gravitational instability, slid toward the southeast and pushed up the arcuate wrinkle ridges (Figure 10). In this case one would expect asymmetrical cross-sections of the ridges with the flatter side oriented toward the summit and the steeper side of the ridges oriented away from the summit. MOLA data seem to support this (e.g., Figure 2a). Crumpler et al. [1996] reported that gravity-induced flank structures may be divided into stresses that are related to (1) the relief of the volcano, (2) the regional slopes on which the volcanoes are constructed, and (3) the loading of the lithosphere by the mass of the volcano. They interpreted mare-type wrinkle ridges on the upper flanks of Asraeus Mons as products of lithospheric loading. Alternatively, magma chamber inflation resulting in uplift of parts of the volcano can cause steepening and sliding of the volcano flanks and wrinkle ridge or terrace formation [Crumpler et al., 1996]. Yet another alternative, completely independent of volcanic/tectonic processes is that the rims of large impact craters that were buried with younger Syrtis Major lavas could produce the observed ridge pattern. In this case we would expect to find arcuate ridges everywhere along the margins of Syrtis Major where the lavas are thinner. This is not consistent with our observations (Figure 9).

However, there is good evidence that Syrtis lavas flooded older craters and that at least some of the wrinkle ridges are related to, and located above, completely buried impact craters (Figure 9c). This third type of wrinkle ridges is more or less circular and continuous in map view. These wrinkle ridges occur on the entire shield except the very northwest. Although statistics are poor there seem to be more and larger circular wrinkle ridges in the eastern and southwestern regions of Syrtis Major. This could be related to thinner lava thicknesses in these areas allowing older, buried crater rims to be still visible in form of circular wrinkle ridges. In areas with greater lava thickness, old craters might be buried so deeply that there is no evidence of their existence left on the surface. Provided these wrinkle ridges are indeed buried impact craters, we can estimate the thickness of the covering layer by using the relationships between diameters and rim heights of fresh impact craters [e.g., DeHon and Waskom, 1976; Hiesinger et al., 2002]. The diameters of completely closed circular wrinkle ridges that we investigated in detail vary from ~5 to ~60 km. The corresponding rim height can be calculated with power laws derived from a global study of Martian impact crater morphologies by Garvin et al. [2002]. They found that the rim height of fresh impact craters smaller than 7 km in diameter is given by \( h = 0.07D^{0.52} \), with \( D \) being the crater diameter. Rim heights of craters between 7 and 110 km are given by \( h = 0.05D^{0.60} \) [Garvin et al., 2002]. On the basis of these relationships between diameters and rim heights of craters we estimated the thickness of the Syrtis lavas to be less than ~600 m, a thickness which is in excellent agreement with estimates based on MOLA profiles (section 3.5).

### 3.8. Calderas

Crumpler et al. [1996] proposed that gravity-induced flank structures may be divided into stresses that are related to (1) the relief of the volcano, (2) the regional slopes on which the volcanoes are constructed, and (3) the loading of the lithosphere by the mass of the volcano. They interpreted mare-type wrinkle ridges on the upper flanks of Asraeus Mons as products of lithospheric loading. Alternatively, magma chamber inflation resulting in uplift of parts of the volcano can cause steepening and sliding of the volcano flanks and wrinkle ridge or terrace formation [Crumpler et al., 1996]. Yet another alternative, completely independent of volcanic/tectonic processes is that the rims of large impact craters that were buried with younger Syrtis Major lavas could produce the observed ridge pattern. In this case we would expect to find arcuate ridges everywhere along the margins of Syrtis Major where the lavas are thinner. This is not consistent with our observations (Figure 9).

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On the basis of the arcuate orientation of three sets of concentric graben, Schaber [1982] proposed a 280 km wide
circular depression that encloses Nili Patera (caldera C$_1$ of Schaber [1982]) and Meroe Patera (caldera C$_2$ of Schaber [1982]). However, new MOLA data indicate that Nili Patera and Meroe Patera are located within a complex larger (350 × 150 km) N-S elongated depression rather than a circular depression (Figure 11). The highest point of Syrtis Major is NW of Nili Patera at ~2300 m. The terrain southwest, west and north of the calderas is noticeably
Figure 11. Detailed view of the MOLA topography associated with Nili and Meroe Patera. The calderas are located within a larger N-S elongated central depression. Slopes of western Nili Patera and eastern Meroe Patera are much steeper and height differences to the surrounding terrain are much larger compared to the opposite sides of each caldera. Contour lines are spaced 100 m apart.
higher (~2000–2300 m) than east and southeast of the calderas (~1500–1700 m) and forms a crescent-like summit with steeper slopes into the summit depression and gentler slopes away from the summit. The caldera floors are at an elevation of ~100–200 m and 180–250 m, respectively. MOLA data indicate that the floor of Nili and Meroe Patera are ~1 km below the inferred pre-volcanic terrain and at about the same elevation as the cratered highlands immediately north of Syrtis Major (Figure 6). This is a major difference to most other Martian volcanoes (e.g., Olympus, Ascraeus, Arisia, Pavonis and Elysium Mons, Alba Patera) with their caldera floors at much higher elevation than the surrounding terrain. Wilson and Head [1994] showed that secondary magma reservoirs form at levels where ascending magma reaches neutral buoyancy. With the large heights of the volcanoes mentioned before, this zone of neutral buoyancy might be located within the edifice. Drainage or collapse of such shallow reservoirs cause the formation of the calderas, dike propagation and large-scale flank eruptions (e.g., Arisia Mons). The implication for Syrtis Major is that its magma reservoir is still in the megaregolith and not within the volcanic construct as in the case of the large Tharsis volcanoes and Elysium Mons. Support for such an interpretation comes from terrestrial analogs. “Shallow” depths to magma chambers on Earth are on the order of several kilometers (e.g., 4–7 km for the Long Valley caldera (USA), 2–4 km for Rabaul caldera (New Guinea) [Lipman, 2000], a depth considerably deeper than the height of the entire Syrtis Major volcanic construct. Rim monoclines associated with granodiorite and diorite plutons in the Great Basin of western North America, which suggest relative downward movement of the pluton while wall rocks were hot and ductile, led Glazner and Miller [1997] to propose that the plutons rose to a level of neutral buoyancy and then sank as their densities increased during crystallization. Such a downward movement of a relatively dense pluton could cause pluton-down faults and grabens along their margins and surface depressions above the sinking pluton [Glazner and Miller, 1997].

[Crumpler et al. [1996] investigated Martian calderas in great detail. According to their work, Nili and Meroe Patera are ~0.5 km deep. It is not quite clear what base level Crumpler et al. [1996] used to estimate the depth of the calderas but MOLA data confirm that Nili and Meroe Patera are ~0.5 km deeper than the floor of the large central depression (Figure 11). However, height differences between the caldera floor and the surrounding terrain on the western edge of Nili Patera and the eastern edge of Meroe Patera are much larger and are on the order of ~1.5 km and ~1.0 km, respectively. Table 2 of Crumpler et al. [1996] is a summary of morphological characteristics of Martian calderas. On the basis of their compilation of data, Nili and Meroe Patera appear unusual because they do not share principal geological characteristics with other calderas on Mars. Compared with other Martian calderas, Nili and Meroe Patera do not have terraced or furrowed caldera margins, circumferential fracture or ridge patterns or radial fracture/ridge patterns. In addition, no nested sequence of calderas or overlapping sequence of calderas has been observed for the calderas of Syrtis Major. Crumpler et al. [1996] also found no evidence for linear or arcuate fissure eruptions, parasitic cones, radial or fan-shaped flank patterns, flank pits/rilles or flank terraces. However, Crumpler et al. [1996] described Nili Patera as a 40–70 km diameter C-shaped asymmetric depression

Figure 12. Detailed views of (a) Nili Patera and (b) Meroe Patera based on a Viking MDIM image mosaic. Nili Patera exhibits numerous graben and normal faults on its west side, a semi-circular ridge and a straight fissure or graben on the floor as well as a cone-like feature and bright areas. Meroe Patera is characterized by its smooth floor, fewer and less prominent graben and faults, no albedo differences, and a higher crater density. 

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that marks the location of significant fissure eruptions. They described a pattern of parallel graben and faults at the western edge, a central NNE-SSW graben and a small breached cone in the northeast of the smooth floor, which is indistinguishable in surface texture and albedo from the surrounding plains. Meroe Patera is \( 40 \) km in diameter, more circular in outline, and defined by a near continuous caldera scarp [Crumpler et al., 1996]. Crumpler et al. [1996] also found an “unusual castellated texture of ridges” that may represent the surface of a late-stage lava lake or eolian material in Meroe Patera. They interpreted the absence of mare-type ridges as evidence for either the latest stage of subsidence post-dating the formation of the mare-type ridges or for unfavorable mechanical characteristics of the caldera floor material that hindered ridge propagation across the floor.

[27] On the basis of our inspection of Viking MDIM data, we make the following observations. The western and northern margins of Nili Patera are characterized by several graben and normal faults and a steep cliff-like scarp. Such a scarp is absent in the east and southeast of Nili Patera (Figure 12a). The western floor of Nili Patera shows a semi-circular ridge with divergent and convergent ridge segments in the south. Between the ridge and the western cliff we observe a rough terrain, which might have been formed by mass wasting along the cliff. Central parts of the smooth floor of Nili Patera are characterized by a roughly north-south trending fissure or graben and a bright albedo feature. In the northeast we observed a cone-like feature (Figure 12a). Meroe Patera is morphologically different to Nili Patera. Graben associated with Meroe Patera are less prominent, smaller in size, and smaller in

![Figure 13. Detailed views of Nili Patera based on THEMIS day- and nighttime IR images.](image-url)

(a) Part of daytime image I01508007 of the western edge of Nili Patera. Numerous graben and normal faults are visible, as well as a ridge on the caldera floor. In the northern parts of Nili Patera we observe albedo differences, which we interpret as layering within the wall. (b) Part of nighttime image I01627002 of the western edge of Nili Patera. Northern parts appear bright, indicating coarse grained material or bedrock. (c) Part of daytime image I02207005 of the eastern edge of Nili Patera showing a cone-like feature, a lobate low albedo area and an area with a leather-like surface texture. (d) Part of nighttime image I00903005 of the eastern edge of Nili Patera. In this image, the low albedo area of image I02207005 appears bright, indicating coarse grained material or bedrock. The terrain with the leather-like surface texture appears dark, indicating finer grain sizes. The scale bar for this image is missing because there are no geometric data available.
number. We do not observe ridges, graben or cone-like features on the smooth floor of Meroe Patera and wrinkle ridges that occur outside the caldera stop right at the caldera margins (Figure 12b). Meroe Patera is characterized by a semi-circular steep cliff-like scarp in the east, which is absent in the west. Instead of rough terrain that we observed along the scarp of Nili Patera, we observe a terrace-like feature along the scarp of Meroe Patera. A prominent albedo feature is absent on the floor of Meroe Patera but we see two sinuous lava channels in the south and southeast of Meroe Patera. Another difference between Nili and Meroe Patera is the number of superimposed impact craters, with Nili Patera showing fewer craters than Meroe Patera (Figure 12b).

[25] Daytime THEMIS IR image I01508007 covers the western part of Nili Patera and shows that the rough terrain between the ridge and the scarp consists of numerous knobs. In this image we also see fine-scale albedo differences in the inner caldera wall, which we interpret as layering (Figure 13a). Nighttime THEMIS IR image I01627002 shows bright spots in the northern parts of the floor of Nili Patera, which we interpret to indicate bedrock or blocks that are not covered with fine-grained material (Figure 13b). Daytime THEMIS IR image I02207005 covers the eastern part of Nili Patera. In this image we see that the cone-like feature has steep slopes and probably a central depression. Southeast of the cone we observe a low albedo terrain and south of the cone we see a rough bright terrain with a leather-like texture that might be caused by small-scale dunes. This terrain appears to correspond to the bright terrain identified in Viking MDIM data (Figure 13c). In the MDIM data, the dark terrain southeast of the cone appears only marginally darker than the surrounding terrain and based on these data one would probably not map this terrain as a separate unit as is suggested by the THEMIS data. Nighttime THEMIS IR image I00903005 indicates that the dark area southeast of the cone consists of material which retains heat well, that is coarse-grained material. In this image the bright terrain with leather-like texture appears dark, indicating fine-grained material (Figure 13d). Daytime THEMIS IR image I01096005 shows very subtle polygonal patterns, possibly ridges, on the floor of Meroe Patera as well as several areas with dark albedo. This image also confirms that Meroe Patera is covered with a much larger number of craters than Nili Patera (Figure 14a). Nighttime THEMIS IR image I00878002 of Meroe Patera exhibits a semicircular ring of bright terrain on the western central floor, which reflects a more coarse-grained nature of the material compared to the rest of the floor. The polygonal terrain may also be outlined by somewhat coarser-grained material, which probably forms subtle ridges (Figure 14b).

[29] On the basis of Magellan radar data, Crumpler et al. [1997] investigated 96 calderas larger than 20 km on Venus, which are characterized by subsidence in association with eruption of volcanic material at the surface. Compared to Martian calderas we recognized morphologic differences and similarities. Numerous Venusian calderas exhibit a complex concentric pattern of fractures, frequently so dense in spacing that the character of lava flows is obscured near the caldera margins [Crumpler et al., 1997]. Crumpler et al. [1997] concluded that, as is the case with Arsia Mons, some of the Venusian calderas are comparable to down-sag calderas known from Earth [e.g., Walker, 1984]. Many calderas on Venus are characterized by flat floors and steep walls, are relatively deep by terrestrial standards (several kilometers) and are highly circular in plan view with respect to the often irregular, frequently overlapping, and generally flat-floored morphology of terrestrial or Martian calderas. In terms of size and morphology Venusian calderas are similar to the Arsia-type calderas on Mars [Crumpler et al., 1997]. However, the caldera of Pavonis Mons, which is ~5 km deep, is the only Martian caldera that is comparable in depth to calderas on Venus and might have formed by drainage of the underlying reservoir during voluminous flank eruptions. The greater depth and circularity of Venusian calderas could be related to relatively greater predicted depths of the reservoirs [Crumpler et al., 1997].

3.9. Flooding of Crater Antoniadi, Nili Fossae, and an Occurrence of Unit Hs West of Syrtis Major

[36] The geologic map of Meyer and Grolier [1977] shows two separate units within crater Antoniadi. Unit pr consists of ridged plains, forms an anulus along the lower inner crater walls and also occurs on Syrtis Major. A smooth plains unit (p), which postdates the Syrtis Major lavas (pr) is exposed in the center of crater Antoniadi. The more recent geologic map of Greeley and Guest [1987] indicates that crater Antoniadi is flooded with lavas from Syrtis Major (Figure 1). We used MOLA and MOC data to investigate this in detail. MOLA data show that there is only a very narrow passage into crater Antoniadi and any lava from Syrtis Major would have to flow through this spillway in order to fill Antoniadi (Figure 15a). Wide-angle MOC image M0305716 confirms our observation that there is only a small passage into crater Antoniadi (Figure 15b). This image shows a very sharp contact between lava and pre-existing highland terrain and a crater rim. We interpret this as clear evidence for lava embaying older, topographically higher terrain that has not been covered with Syrtis Major lavas. On the basis of MOC image M0305716 the width of the passage into crater Antoniadi is on the order of 5.5 km.

[31] In addition, MOLA data show that the floor of crater Antoniadi is higher toward the north, so lava would have to run-uphill over a distance of ~250 km in order to reach the extent of unit Hs as mapped by Greeley and Guest [1987]. Figure 16 shows a profile across crater Antoniadi, which indicates that the northern part of the crater floor slopes toward the south with ~0.06 degree. One might argue that this upsillslope was formed after the emplacement of the lava by subsequent subsidence of the crater floor due to loading with lava. If the amplitude of subsidence and tilting were large, we would expect numerous wrinkle ridges in crater Antoniadi due to the compressional stress regime associated with subsidence of a crater/basin floor. However, careful inspection of MOLA data and imaging data revealed only a small number of wrinkle ridges within the crater. On the basis of our observations we propose that comparatively minor amounts of lava flowed through the spillway and that most of the lavas in Antoniadi very likely erupted from sources within the crater.

[32] Nili Fossae is a large graben at the northern edge of Syrtis Major that disrupts Noachian etched terrain and Noachian plains (units Nple and Npl2 of Greeley and Guest
Figure 14. Detailed views of Meroe Patera based on THEMIS day- and nighttime IR images. (a) Part of daytime image I01096005 of the eastern edge of Meroe Patera. Note the dark albedo pattern that has not been observed in Viking data, the large number of impact craters, and the small-scale ridges on the floor of Meroe Patera. (b) Part of nighttime image I00878002 of the eastern edge of Nili Patera. Dark material of image I01096005 appears to consist of coarse grained material or bedrock. The ejecta of the larger crater in the northeastern floor seems to be blockier than the surrounding area. Also note the ridge pattern of the floor of Meroe Patera.
Figure 15. (a) Detailed view of the MOLA topography associated with crater Antoniadi. White arrows outline the continuous rim of crater Antoniadi that has not been covered with Syrtis Major lavas. Black and white arrow shows the location where the rim has been breached by lavas. White box indicates the location of wide-angle MOC image M0305716. Scale bar is 100 km. Contour lines are spaced 100 m apart. (b) MOC image M0305716 shows a narrow passage into crater Antoniadi (white arrow). Scale bar is 25 km.
Greeley and Guest [1987] mapped the floor of Nili Fossae as lightly cratered, featureless, flat plains (unit Aps) that were probably formed by eolian processes (Figure 1). According to their map Syrtis Major lavas did not enter the graben or at least are now covered by unit Aps. On the basis of the shape of the geologic contact between unit Aps and unit Hs it appears that the graben was thought to be higher in elevation than the front of the volcanic flows of unit Hs,

Figure 16. MOLA gridded topography and a profile across the floor of crater Antoniadi. The upper part of the figure shows the location of data points of the profile in the lower part of the figure. The profile starts at the white point in the southeast and follows the data points toward the northwest. Note that the northern floor of Antoniadi is tilted to the south with ~0.06°. Vertical exaggeration of the profile is ~90x.
hence not permitting lava to flow through the graben. This interpretation is not supported by MOLA data that indicate that Nili Fossae slopes to the north with ~0.05 degree and exhibits a smooth flat floor. The floor of Nili Fossae at its southern end at ~19° N is at about ~400 m elevation and drops to ~1200 m at 27° N, where Nili Fossae crosses the dichotomy boundary (Figure 17). At the dichotomy boundary we observe a break in slope. North of the dichotomy boundary Nili Fossae slopes with ~0.34 degree toward north (Figure 18). At the northern end of Nili Fossae we also observe a lobate feature that might have been created by lava or any other material that flowed through Nili Fossae (Figure 17). As indicated by the bulging of the contour lines in Figure 17, the lobate feature can be traced to ~1900 m. On the basis of this finding the estimated dimensions of the lobe are ~80 × 100 km.

[33] There are 12 MOC images available for Nili Fossae to test this idea. Images of the floor of Nili Fossae generally show identical terrain types that have also been identified elsewhere on the Syrtis Major complex (see section 3.12), that is, cratered terrain and etched terrain. Three MOC images (M0900436, M0902300, M1102709) show evidence of morphologic features that we interpret to be flow fronts (Figure 19). One of these images (Figure 19a) is located at the northern end of Nili Fossae (M0900436), one (Figure 19b) is located even farther northeast on the smooth surface of the lobate feature that we described above (M0902300), and one image (Figure 19c) is located in central Nili Fossae (M1102709). MOC image M0900436 shows layering in the walls of Nili Fossae but it remains ambiguous whether this layering is volcanic or sedimentary in origin. On the basis of MOLA and MOC data we conclude that the slope of Nili Fossae (~0.05°), the smooth flat floor, the lobate feature at the northern end of the graben, the identification of possible flow fronts, and the identified terrain types are consistent with the hypothesis that Syrtis Major lavas flowed through Nili Fossae toward the northern lowlands.

[34] The geologic map of Greeley and Guest [1987] shows a small isolated occurrence of unit Hs just west of Syrtis Major that is located within two craters of ~100 km diameter (300°W, 7.5°N). In their map this occurrence of unit Hs is surrounded by Noachian etched terrain (Nple) and a possible connection to the main body of unit Hs appears to be obscured by two young superposed impact craters (Figure 1). New MOLA data indicate that north of these two young craters, there is a breach in the rim of one of the large craters that connects the two occurrences of unit Hs (Figure 20). The floors of the two 100 km craters are tilted to the south; the southern parts being ~150–200 m lower in elevation than the northern parts close to the spillway. This suggests that lavas that entered the two large craters through the spillway flowed toward the south, following the local slopes within the craters. From our observations we conclude that lavas within the 100 km craters are connected with the Syrtis Major flows and were emplaced contemporaneously.

3.10. Isidis Rim

[35] Syrtis Major is located in close proximity to the Noachian age Isidis impact basin (Figures 1 and 2a). The rim of this basin is unusually low in elevation in several locations such as the passage to the northeast into Utopia/Elrysium Planitia. Rim massifs are also missing where Syrtis Major lavas flowed into the Isidis basin. However, Isidis is well defined along its southern edge by the Libya Montes, as well as in the northeast of the Syrtis Major structure. In the geologic map of Greeley and Guest [1987] the southern rim (i.e., Libya Montes) consists of rough, hilly, fractured material (unit Nphl) of moderately high relief which they interpreted as ancient highland rocks and impact breccias that were formed during the period of heavy bombardment. The units that are exposed at the north rim (Nple, Npl2) are younger than the units of the south rim and are characterized by eolian dissection, collapse of ground ice, minor fluvial activity, and thin lava flows or sediments that partly bury underlying rocks [Greeley and Guest, 1987].

[36] Figure 21 contains a profile along the outer ring structure of the Isidis basin of Schultz and Frey [1990]. On the basis of this profile we see differences in elevation of the investigated ring of more than 7000 m. Highest elevations (~3400 m) are associated with isolated peaks of the Libya Montes; lowest elevations (~3600 m) occur in the passage to the Utopia basin and the northern lowlands. The Libya Montes are characterized by a rough topographic profile with large variations in elevation (~4800 m), whereas the rim segments in the northeast and in the area of Syrtis Major are much smoother at the sampling resolution. The variations in elevation of the Isidis rim can have several causes such as initial heterogeneities, erosion of parts of the rim, tectonic deformation or burial with Syrtis Major lavas [e.g., Wichmann and Schultz, 1988; Tanaka et al., 2002]. Despite the wide range of possible causes, the absence of the Isidis rim has been used as an argument for the emplacement of thick (1–2.5 km) Syrtis Major lavas that covered the rim [e.g., Wichmann and Schultz, 1988]. In this section we will discuss different models to explain the presently observed topography. A very fundamental observation we can make is that even in areas that are not influenced by Syrtis Major lavas there is no “typical” Isidis rim. Instead, the rim morphology is highly variable, ranging from steep isolated peaks in the south, to horst-like areas formed by numerous graben in the north and southeast, to very low lying smooth regions in the northeast. For our discussion of the Isidis rim we will differentiate between the “elevation” of the rim and the “height” of the rim; “elevation” relating to the absolute topography, and “height” relating to the differences between rim and the surrounding terrain.

[37] Figure 22 shows several models of the evolution of the Isidis rim. There is a model that calls for initial heterogeneities of the rim topography to explain the present-day topography, that is, if the Isidis rim in the Syrtis Major area was never as high as in other parts, it could be buried with Syrtis Major lavas. Alternatively, we discuss models that involve erosional or tectonic lowering of parts of the Isidis rim prior to or contemporaneous with the volcanic activity associated with Syrtis Major. For example, Tanaka et al. [2001a, 2001b; 2002] suggested that magmatic activity in the Syrtis Major region began with shallow sills that drove catastrophic erosion of friable upper crustal Noachian rocks, i.e., the Isidis rim (Figure 22a). They proposed that these rocks were charged with water ice, water, or perhaps CO2 ice or CO2 clathrate, allowing large volumes of rock to be disrupted, eroded and transported for many hundreds
of kilometers (also see Hoffman [2000]; Tanaka et al. [2001]). According to Tanaka et al. [2001a] the eroded material would have ultimately filled the Isidis basin with tens to a few hundreds of meters of sediments. On the basis of TES thermal inertia data Bridges et al. [2003] concluded that most of the rock debris in the Isidis basin came from the southern Noachian highlands and this observation is inconsistent with the majority of rocks being brought in from Syrtis Major as suggested by Tanaka et al. [2000, 2002]. Interestingly, Tanaka et al. [2002] reported that the “topography does not show clear evidence for erosion of the Isidis rim, but the volcanic burial may have been sufficiently thick to largely mask any erosional record”. As a variation of Tanaka’s model, interaction of dikes and lava flows with a volatile-rich substrate could have driven erosion of the Isidis rim (Figure 22b). In this case dikes from beneath and lava flows from above could have provided heat in order to release enough volatiles to collapse the rim. North of Syrtis Major a system of large graben can be observed. Some of these graben, e.g., Nili Fossae (see discussion above), are flooded by Syrtis Major lavas and there is the possibility that these graben originally extended further to the south and were subsequently covered with lava. In this model graben formation could have lowered the Isidis rim before Syrtis Major lavas were emplaced (Figure 22c). Graben formation and contemporaneous eruption of lavas along the newly formed graben also appears plausible. Alternatively, loading with early Syrtis Major lavas could have initiated fault propagation along an already fractured and friable Isidis rim. Loading this area with even more, younger lavas could have destabilized the rim and caused it to collapse into the Isidis basin (Figure 22d). Finally, Syrtis Major lavas might simply be thick enough to cover the Isidis rim, especially if the original rim was at low elevations as a result of initial heterogeneities (Figure 22e).

[35] In order to test these ideas and to see how the present-day topography might have formed, we used MOLA data to simulate the Isidis rim in locations where it has supposedly been eroded, that is, along its western margin. For this purpose we first have to characterize the Isidis rim and know its topography and morphology. MOLA data indicate that the highest peaks of the still preserved Isidis rim occur in the south (Figure 21). These peaks are between 2000 and 2500 m high, with some of them being up to 3500 m in elevation. However, even at the south rim only very isolated peaks exhibit such high elevations, the majority of the terrain is much lower (<2000 m). In addition, the morphology of unit NpH differs significantly between the south and the southeast of the Isidis rim. The southeast is generally much lower in elevation, high peaks are absent, graben are frequent, and the terrain resembles much more the terrain at the north rim in terms of morphology and topography. In the north the rim is characterized by a plateau-like terrain, which is generally lower in elevation with the summits of the highest peaks/plateaus at ~1000–1900 m elevation. These differences in topography and morphology along the rim are primarily interpreted as initial heterogeneities of the Isidis rim that likely also occurred in the Syrtis Major region.

[39] The Isidis basin can be traced across the dichotomy boundary [e.g., Grizzaffi and Schultz, 1989; Schultz and Frey, 1990], which is characterized by topographic differences between southern highlands and northern lowlands of at least 2.5 km and average slopes of ~1° [Frey et al., 1998]. On the basis of MOLA data, Smith et al. [1999a] determined a pole-to-equator upward slope of ~0.036 degrees from the northern lowlands toward the southern highlands at all longitudes. We found similar slopes of ~0.03–0.07 degrees over baselines of ~1500 km across the Isidis basin toward the north. From these observations we conclude that the basin is located on a northward tilted surface. Consequently, it might be an initial characteristic of the basin formation that the north rim is lower in absolute elevation than the south rim. In addition, oblique impact geometry, locally more extensive rim erosion, different styles of erosion, different types of rim material, or variations in tectonic style and amounts of isostatic adjustment might have contributed to the observed topography. For example, MOLA data indicate that the floor of the Isidis basin is tilted toward the southwest with ~0.02 degree. This southwest slope has been attributed to loading of the northern lowlands with sediments of the Vastitas Borealis Formation [Tanaka et al., 2001c] or km-thick ridged plains units [Watters, 2003]. It was proposed that such loading of the northern lowlands could have resulted in extensive deformation of the lithosphere and the tilt of the Isidis floor [Tanaka et al., 2001c; Watters, 2003]. In summary, there is a wide variety of geologic processes that can cause the observed topographical and morphological differences along the Isidis rim. With this in mind, we will now try to deconvolve the complex nature of the Isidis rim.

[40] Figure 2a shows a detailed view of the present-day topography, and the location of rim segments that we used to create two separate simulated continuous Isidis rims. For our rim simulations we chose areas that are part of the actual northern Isidis rim, have not been flooded with lavas, and are located immediately north of the Syrtis Major complex. The location of our test areas along the northern rim was selected because this part is characterized by larger continuous areas of high topography compared to the southeastern regions. The test areas were also chosen to be similar to or higher in elevation than southeastern parts of the rim in order to be as representative of the Isidis rim as possible. In one case the test area is ~0–1200 m in elevation, in the second case the area is ~500–1700 m in elevation (area 1 and 2 in Figure 2a). With these two terrain samples we created artificial Isidis rims along the inferred location of ring structures derived by Schultz and Frey [1990]. With the two created rim segments in place (outlined in white in

Figure 17. Detailed view of the MOLA topography of Nili Fossae and adjacent terrain. The floor of Nili Fossae is tilted to the northeast and shows a drop in elevation of several hundreds of meters along the graben. MOLA data show a lobate delta-like feature at the northeastern mouth of Nili Fossae (white arrows), suggesting that some fluid material, very likely lava, flowed through the graben. Black arrows indicate the location of a steep cliff at the passage into the Isidis basin. Contour lines are spaced 100 m apart.
Figure 18. MOLA gridded topography and a profile along Nili Fossae. The upper part of the figure shows the location of data points of the profile in the lower part of the figure. The profile starts at the white point in the southwest and follows the data points toward the northeast. Nili Fossae slopes with ~0.05° toward the northeast. Slopes increase to ~0.34° where Nili Fossae crosses the dichotomy boundary. Vertical exaggeration of the profile is ~135x.
Figure 2b), we could compare the simulated rim topography to the actual MOLA topography of the same region. Black arrows in Figure 2b indicate locations where the simulated ring topography of scenario 2 (500–1700 m) is higher than the present-day topography. As a result of this investigation we see that our first simulated ring, 0–1200 m in elevation, can be covered to \( \approx 75\% \) with Syrtis Major lavas. The second simulated ring, 500–1700 m in elevation, can be covered to more than 60 percent. From this we conclude that most of the simulated topography that is similar to the northern and southeastern rim of Isidis can be covered with Syrtis Major lavas without any involvement of additional erosion or modification of rim mountains. However, if the northern test areas are representative of the Isidis rim, we should still see isolated massifs protruding through the lavas similar to Figure 2b, even more so had we chosen terrain samples from the Libya Montes for our rim simulation. As we do not observe such massifs one has to conclude that these massifs either never existed in the western parts of Isidis or that one or several geologic processes removed these massifs. For this reason we propose that initial heterogeneities and/or flooding with lavas, together with erosion and/or tectonic modification, contributed to the present-day topography of the Isidis rim.

3.11. Age of Syrtis Major

[41] On the basis of crater densities, TANAKA et al. [1992] proposed a global Martian stratigraphy. According to this stratigraphy, a geologic unit is of Early Noachian age if it has more than 200 craters per \( 10^5 \) km\(^2\) with diameters larger than or equal to 16 km (\( N(16) > 200 \)). Together with the Middle Noachian (\( N(5) > 400 \)) and Late Noachian (\( N(5) = 200–400 \)) this forms the oldest Martian epoch, the Noachian. Younger than the Noachian is the Hesperian, which is subdivided into Early Hesperian (\( N(5) = 125–200 \)) and Late Hesperian (\( N(5) = 67–125 \)). The youngest epoch is the Amazonian, which is subdivided into Early Amazonian (\( N(2) = 150–400 \)), Middle Amazonian (\( N(2) = 40–150 \)), and Late Amazonian (\( N(2) < 40 \)).

[42] In the geologic map of GREELEY and GUEST [1987], Syrtis Major is represented by unit Hs which is of middle to Late Hesperian age (\( N(5) = \approx 80–115 \)). In order to investigate/verify the stratigraphic position of the Syrtis Major lavas in more detail, we performed new crater counts for the entire shield. Using Viking MDIM data, we counted more than 4000 craters, shown as white dots in Figure 23. As will be discussed later (section 3.12), MOC images show evidence for erosional modification of Syrtis Major. Craters observed in MOC images are mostly below the image resolution of Viking MDIM data on which we performed our crater counts. On the basis of our measured crater size frequency distribution, we see that craters smaller than \( \approx 1 \) km diameter are underrepresented on the counted area (Figure 24). This can have two reasons. First, poor image resolution makes it impossible to identify smaller craters. Second, small craters might have been destroyed by geologic processes. Such a destruction of small craters could be caused by erosion of the terrain as observed in the MOC images. This effect is largest or most effective at small crater sizes and will only to a lesser degree influence the frequency distribution of larger craters that were used to derive the age of Syrtis Major.

[43] If we apply the TANAKA et al. [1992] stratigraphy, our new crater counts indicate that Syrtis Major is of Early Hesperian age (Figure 24). We find that the total number of craters with \( D \geq 5 \) km per \( 10^5 \) km\(^2\) is \( \approx 154 \), hence Syrtis Major is slightly older than mapped by GREELEY and GUEST [1987]. The statistical error of our crater counts as given by

\[
\Delta N = \sqrt{N}
\]
Crater ages of $N(16) = 20 \pm 2.2$ also indicate an Early Hesperian age, whereas crater ages of $N(2) = 679 \pm 13$ indicate a Late Hesperian age (Figure 24). However, additional support for an Early Hesperian age of Syrtis Major comes from several other crater count studies [e.g., Condit, 1978; Maxwell and McGill, 1988]. Condit [1978] performed crater counts for intermediate size craters (4–10 km in diameter) on several geologic units, including four ridged plains units. On average, he identified $128 \pm 25$ craters per $10^6$ km$^2$ on the ridged plains, with Syrtis Major showing 141 craters per $10^6$ km$^2$. According to his Figure 7, this corresponds to an Early to middle Hesperian age. In addition, Maxwell and McGill [1988] performed crater counts for Syrtis Major and Isidis Planitia. They reported that the number of craters of a specific diameter can either be read directly from the cumulative crater size-frequency plot, or more commonly, be determined from fitting a production curve to the data. Maxwell and McGill [1988] fitted their counts with the Neukum [1983] production function, but also gave crater ages based on where crater

**Figure 20.** Detailed view of the MOLA topography of two craters at the western edge of unit Hs. The interiors of these craters have been mapped as an isolated occurrence of unit Hs and there is no connection to unit Hs of Syrtis Major in the geologic map of Greeley and Guest [1987]. MOLA data show a narrow passage through which lavas from Syrtis Major could have entered the craters. Contour lines are spaced 100 m apart.
Figure 21. MOLA gridded topography and a profile along the outer Isidis ring of Schultz and Frey [1990]. The upper part of the figure shows the location of data points of the profile in the lower part of the figure. The profile starts in the east and follows the data points clockwise around the basin. Note the low topography of the northeastern ring segment and its smooth appearance. Syrtis Major appears also much smoother than the rough Libya Montes. Libya Montes are comprised of isolated peaks and deep valleys and show a wide variation in elevation of several thousands of meters. Vertical exaggeration of the profile is $\sim 250x$. 
Figure 22. Interpretative diagram (cross-sections), illustrating the proposed models for the evolution of the Syrtis Major-Isidis region. (a) Catastrophic release of volatiles due to sill emplacement causes rim erosion; (b) Catastrophic release of volatiles due to dike and lava emplacement causes rim erosion; (c) Rim collapse due to formation of graben; (d) Loading with early Syrtis Major lavas; (e) Emplacement of Syrtis Major lavas that are thick enough to cover the rim massifs.

distributions cross a given diameter [e.g., Tanaka, 1986]. On the basis of their crater counts, Maxwell and McGill [1988] reported 150 craters larger than or equal to 5 km when fitted to the Neukum [1983] production function and 190 craters larger than or equal to 5 km when directly read from the data [Tanaka, 1986]. On the basis of the Tanaka et al. [1992] stratigraphy, these numbers indicate an Early Hesperian age.

[44] As described earlier (section 3.6), spectral data suggest compositional differences between the eastern and western parts of Syrtis Major [Mustard et al., 1990]. The surface roughness map of Kreslavsky and Head [2000], as well as the MOLA topography and the distribution of large-scale wind-streaks also indicate differences between the east and west of Syrtis Major. For this reason we performed separate crater counts for the eastern and the western parts of Syrtis Major as well as for the floor of crater Antoniadi. Figure 23 shows the boundaries of these crater count areas, which were mapped based on differences in surface roughness. As a result we found $163 \pm 6.4$ craters with diameters larger than 5 km per $10^6$ km$^2$ ($N(5) = 163 \pm 6.4$) on the floor of crater Antoniadi. Western parts of Syrtis Major are slightly younger and exhibit $158 \pm 6.3$ craters with diameters larger than 5 km per $10^6$ km$^2$ ($N(5) = 158 \pm 6.3$). Even younger are the eastern regions of Syrtis Major, which have crater ages of $N(5) = 133 \pm 5.7$. In summary, our crater counts reflect subtle differences between the eastern and western regions of Syrtis Major, as suggested by spectral data, topography, surface roughness and the distribution of large-scale wind streaks. Despite the differences in age, all parts of Syrtis Major are of Early Hesperian age according to the definition of the Martian stratigraphic systems of Tanaka et al. [1992].

[45] Presenting our crater counts for entire Syrtis Major in a cumulative plot [Arvidson et al., 1979], we see that the cumulative crater size-frequency distribution of Syrtis Major is characterized by breaks in slope (Figure 24). Such breaks in slope have been interpreted as an indication of
resurfacing processes such as volcanic flooding, emplacement of thin ejecta blankets, or eolian manteling [e.g., Neukum and Horn, 1976]. Such processes completely or partially destroy impact craters and “reset the clock”. The diameters at which these breaks in slope occur can be used to determine the time of resurfacing. On the basis of our crater counts, we see evidence for a resurfacing event in the Late Noachian-Early Hesperian, suggesting a prolonged time of volcanic activity that led to the emplacement of unit Hs.

[46] Tanaka [1986] and Tanaka et al. [1992] correlated Martian epochs with absolute ages. In their stratigraphy the Noachian lasts from 4.6 to 3.5 b.y., the Hesperian from 3.5 to 1.8 b.y and the Amazonian from 1.8 b.y. to present (Figure 25). According to the Neukum and Wise [1976] stratigraphy the Noachian ranges from 4.6 to 3.8 b.y., the Hesperian from 3.8 to 3.55 b.y. and the Amazonian from 3.55 b.y to present (Figure 25). The largest disagreement between these two stratigraphies was for the Hesperian epoch, being either <300 m.y. long [Neukum and Wise, 1976] or ~1.7 b.y. long [Hartmann et al., 1981; Tanaka, 1986]. Very recently this wide range in age of the Hesperian epoch has been narrowed, the Hesperian now being considered to last from 2.9 to 3.7 b.y. [Hartmann and Neukum, 2001], with Neukum favoring a somewhat shorter Hesperian Period (3.3–3.7 b.y.) compared to Hartmann (2.9–3.5 b.y.). In order to verify our crater counts of Syrtis Major and to derive an absolute model age, we performed independent crater counts with the stereo comparator at the DLR Institute of Space Sensor Technology and Planetary Exploration. These crater counts indicate a model age of ~3.6 b.y. which is, according to the new chronology of Neukum [Hartmann and Neukum, 2001], Early Hesperian, hence consistent with our other counts and counts of Condit [1978] and Maxwell and McGill [1988]. In the new Hartmann chronology [Hartmann and Neukum, 2001], 3.6 b.y. corresponds to a Late Noachian age (Figure 25). No matter which chronology we use, the age of Syrtis Major is older than in the geologic map of Greeley and Guest [1987]. On the basis of this second set of independent crater size-frequency measurements, we again found characteristic breaks in slope of the crater size-frequency distribution, which we interpret as evidence for a resurfacing of Syrtis Major. Such a resurfacing event could either be erosional in nature, that is the destruction of craters, or depositional, that is the covering of craters with lava flows or eolian sediments. On the basis of the Neukum chronology we

Figure 23. Viking MDIM of Syrtis Major (dark area). White filled circles represent the location and size of primary impact craters of the measured crater size-frequency distribution shown in Figure 24.
3.12. Description of Syrtis Major Terrain Types

After Tanaka et al., [1992], we have evidence that the resurfacing event occurred at ~3.7–3.8 b.y. ago, or in the Late Noachian/Early Hesperian.

3.12. Description of Syrtis Major Terrain Types

High-resolution MOC images enable us to investigate the morphology of Syrtis Major in unprecedented detail. Compared to the majority of Viking imaging data (i.e., the Viking MDIM data), the resolution of MOC images is ~50 times better, allowing one for the first time to identify and describe terrain morphologies that were not visible in Viking data. On the basis of MOC images we can address the questions of the origin of major morphological features, the origin of unusual surface roughnesses and the level of surface modifications. As of February 2003 ~200–300 MOC images are available for the Syrtis Major volcanic complex. Most of the images we studied have resolutions between 3 and 10 m per pixel have been map-projected by the MOC team and are available online (http://msss.com/mars_images/). On the basis of our systematic inspection of these images we identified distinctive morphologic features such as craters, scarps, rilles, ridges, hills, knobs, and debris aprons. In addition, evidence for layering was found in numerous locations. We also identified several terrain types with distinctive surface morphologies (i.e., cratered, grooved, etched, pitted, polygonal) (Figure 26). Dark material with locally restricted spatial extent (i.e., dark halo craters, dark eolian deposits) occurs frequently, as well as wind streaks and dunes. Interestingly, we only very rarely found evidence for features that could clearly be interpreted as lava flow fronts.

[s7] Craters were seen in all investigated images, but the density of these small craters is widely variable. In some images craters are rare, in other images the terrain is saturated with small-scale impact craters. Layering, rilles, hills, knobs, and scarps were observed in about every other image and the areal distribution of these features on Syrtis Major appears to be to a first order homogeneous. Debris aprons are associated with steep crater walls, scarps, rilles, hills, and knobs and were found in more than 50% of all studied images. Small-scale wind streaks and dunes are also common and were observed in almost every other image. Looking at the areal distribution of small-scale wind streaks and dunes, we see that these eolian features occur preferentially in the eastern and southern parts of Syrtis Major and within crater Antoniadi. MOC images of the Northwestern quadrangle show, if at all, significantly less small-scale wind streaks and dunes than images of other areas of Syrtis Major. However, as MOC images are not evenly distributed over the entire Syrtis Major complex, the impression that small-scale wind streaks and dunes are scarce in the northwest might be influenced by the uneven sampling density. Large-scale wind streaks as visible in the Viking images show a different and almost opposite distribution. On the basis of the Viking MDIM mosaic it appears that large-scale wind streaks and dunes are more common on the western half (i.e., west of Arena Rupes) and less common or absent in the eastern half of Syrtis Major.

[s8] MOC images of Syrtis Major show several distinctive terrain types (Figure 26). The most frequently observed terrain type is cratered terrain, which is characterized by a very large number of impact craters of a variety of sizes and degradation stages (Figure 26a). These craters exhibit raised rims but no ejecta blankets or secondary craters, and are often superposed on each other. None of the craters shown in Figure 26a shows a fresh crater morphology with well defined continuous and/or discontinuous ejecta blankets. The craters appear to be rather shallow with their floors only slightly below the surrounding terrain. On the basis of these observations and the lack of evidence for material to cover the ejecta blankets, we propose that these craters were exhumed which modified their morphology. The saturation with impact craters and the destruction of older craters by younger craters cause the terrain to be relatively rough. However, in numerous cases younger eolian mantling and dunes cover cratered terrain and have smoothed its morphology (Figure 26b). The size of most of these craters is well below the image resolution of Viking MDIM data. A second cratered terrain unit appears exhumed, and exhibits a rough surface texture that is caused by heavily degraded craters of variable dimensions (Figure 26c). These craters appear to have rounded rims, eroded ejecta blankets and do not exhibit fresh crater morphologies. The rims of these
craters are partially eroded and their floors appear to be at the same elevation as the surrounding terrain. We do not observe secondary craters in the vicinity of the two larger craters in Figure 26c. Owing to destruction of craters by processes associated with burial and exhumation, the crater density of this cratered terrain appears to be smaller compared to the cratered terrain described above. Parts of this terrain appear to have been mantled with a more recent thin veneer of eolian deposits.

Grooved terrain is the second most frequent terrain in Syrtis Major (Figure 26d). This terrain shows roughly parallel oriented ridges and irregularly shaped elongated depressions between the ridges. Some of these ridges and depressions might have formed from crater chains or heavily cratered terrain when parts of coalescent crater rims were eroded. Alternatively, an interpretation as small-scale wrinkle ridges formed by upper boundary layer modifications of a flow seems plausible. Etched terrain is characterized by its rough surface texture with numerous irregularly shaped rimless depressions (Figure 26e). On the terrain around the depressions impact craters are in some cases still recognizable but most of them have been heavily degraded. Frequently crater rims are at least partly destroyed producing an etched-looking, rough texture with small knobs and depressions. Pitted terrain exhibits small irregular or circular depressions, which usually do not have raised rims (Figure 26f). Compared to cratered terrain, pitted terrain appears to be smoother.

Knobby terrain is formed by isolated, rounded, and mostly circular knobs varying in size from <10 to more than hundreds of meters in diameter. The terrain between the knobs appears to be smooth (Figure 26g). The size of these

knobs is below the resolution of the Viking MDIM data, indicating that they are much smaller than the knobs on the floor of the Isidis basin. The knobs of Syrtis Major also differ in their morphology from the knobs in the Isidis basin. In the Isidis basin these knobs are commonly arranged in arcuate chains with pits and elongated depressions superposed on some of the hillocks [Grizzaffi and Schultz, 1989; Bridges et al., 2003; Hiesinger and Head, 2003]. On Syrtis Major we observed more isolated knobs that do not form any particular pattern and do not have central pit craters. Proposed origins of the cones in the Isidis basin include cinder cones [e.g., Plescia, 1980], tuff cones [Hodges and Moore, 1994], phreatomagmatic explosions forming pseudocraters [e.g., Frey and Jarosewich, 1982], mud volcanism [e.g., Tanaka et al., 2000], and ablation of glacial debris [e.g., Grizzaffi and Schultz, 1989]. A discussion of these models on the basis of MOC data is provided by Bridges et al. [2003], but because the size, morphology, and distribution of knobs of Syrtis Major are different, these models might not be relevant to explain the knobs on Syrtis Major. Rather we propose that these knobs are remnants of an eroded surface layer.

[52] Polygonal terrain is the least frequently observed terrain type of Syrtis Major (Figure 26h). The size of the polygons is commonly <50 m in diameter and the only location where we found this terrain type is within the caldera of Nili Patera where lava very likely ponded and slowly cooled and solidified after the eruptive activity declined. A formation of these polygons by solidification of lava is consistent with their location within the caldera, their symmetry, their thermal inertia characteristics, as well as their similarity to terrestrial analogs. On Earth, tetragonal networks on flow surfaces evolve systematically to hexagonal networks by gradual change of orthogonal to nonorthogonal intersections of ~120° [Aydin and DeGraff, 1988]. For the polygons in Nili Patera we observe both orthogonal and nonorthogonal cracks and THEMIS data suggest that the polygons formed on bedrock. THEMIS and MOC images indicate that the polygonal terrain is covered with dunes. At THEMIS resolution these dunes form a dune field with leather-like texture (Figure 13) and MOC high resolution data show barchan-like dunes (Figure 26h). On the basis of MOC image FHA00451 (Figure 26h) the polygonal terrain seems to be rather flat and smooth and we do not observe raised rims at the edges of the polygons. Polygons have a high albedo and are outlined by darker material which exhibits about the same albedo as the large dunes that cover the polygonal terrain. From this we conclude that polygonal cracks are very likely filled with the same material that forms the dark dunes.

Figure 26. Different terrain types as identified in MOC images of Syrtis Major. (a) M1001512: Cratered terrain; (b) M1700435: Smooth mantle/dunes; (c) M2101306: Degraded cratered terrain; (d) M2300433: Grooved terrain; (e) M0802387: Etched terrain; (f) M0904039: Pitted terrain; (g) M0900138: Knobby terrain; (h) FHA00451: Polygonal terrain. Scale bars are 500 m.
As MOC images do not cover the entire Syrtis Major volcanic complex it is impossible to map distinctive geologic units over large areas at this high resolution. However, we can make observations of the general relative topographic relationships between two neighboring units within each image. By synoptically looking at all available MOC images we can then study whether this specific relative topographic relationship holds for the rest of the available images. Figure 27 shows the different terrain types and how often a certain relative topographic relationship has been observed in the MOC images. As not all available MOC images show clear relative topographic relationships between terrain types, the number of images that can be used for this study is limited to \( \sim 50 \) images. In 27 MOC images of Syrtis Major we found evidence that cratered terrain is covered with a smooth mantle or dunes (e.g., M1700435). In four images (M0202880, M0204423, M0307416, M0402098) cratered terrain is topographically above the pitted terrain, in three images (M0806155, SP126605, SP237606) it is topographically above the knobby terrain and in one image (FHA00451) cratered terrain is topographically above the polygonal terrain and etched terrain (M2202032), respectively. On the basis of the frequency of the observed topographic relationships we conclude that most likely cratered terrain is topographically above the pitted terrain and knobby terrain. Evidence for the topographic relationship between pitted terrain, etched terrain and knobby terrain is less obvious. We found one image (M1300670) that indicates that pitted terrain is topographically above the etched terrain and one image (M0403937) that shows pitted terrain topographically above the knobby terrain. However, we observed in six images (M0305715, M0402099, M0701307, M0900138, M1103908, M1800063) that etched terrain is topographically above the knobby terrain, hence it is very likely that etched terrain occurs topographically between the pitted terrain and the knobby terrain. In two images (M0701307, M0802387) we see that etched terrain is also topographically above the grooved terrain. As no direct contact between grooved terrain and knobby terrain has been observed, the topographic relationship between these two terrain types remains unclear. Finally, one image (M2101306) shows that knobby terrain is topographically above a heavily degraded, exhumed cratered unit.

For a few MOC images that we used for our topographic study, there are MOLA data available that allow us to look into the topographic position of certain terrain types in more detail (e.g., M0202880, M0305715, M0307416, M0802387, FHA00451). These images together with the superposed registered MOLA data are available online at http://pdsimg.jpl.nasa.gov. Only in MOC images M0305715 and M0802387 the MOLA data cross the area with clear stratigraphic relationships. Knobby terrain in MOC image M0305715 occurs on a northward tilted slope and there is a distinctive step in topography of tens of meters at the boundary between knobby terrain and topographically higher etched terrain (Figure 28a). MOLA data indicate that in MOC image M0802387 the etched terrain is \( \sim 50–60 \) m higher than the grooved terrain (Figure 28b). In MOC image M0202880 the MOLA profile does not directly cross the area with the best contact. However, if we trace the cliff that is formed by cratered terrain to the point where MOLA data cross it, it appears that cratered terrain is topographically slightly higher than pitted terrain. In MOC image M0307416 MOLA data do not cross the
Figure 28. Two examples of terrain contacts in MOC images and MOLA data. Images in the center show the location of MOLA measurements superposed on MOC images, and white boxes show locations of enlarged views on the right side. MOLA profile on the left side. (a) Contact between knobby terrain and etched terrain in MOC image M0305715, Scale bar is 500 m; (b) Contact between grooved terrain and etched terrain in MOC image M0802387, Scale bar is 500 m.
contact between cratered terrain and pitted terrain. Topographic differences between cratered terrain and polygonal terrain in MOC image FHA00451 seem to be very small, but again, the area with the best contact is not covered by MOLA data. On the basis of MOLA data and our investigation of geologic units with distinctive surface textures in numerous MOC images we propose the following relative topographic relationships (Figure 29): smooth terrain/dunes (topographically highest), cratered terrain, pitted terrain, etched terrain, knobby terrain, old cratered terrain (topographically lowest). In this scenario, we interpret the cratered terrain to be exhumed from underneath a terrain unit that has been removed and is no longer observable. As described earlier, the interpretation that the cratered terrain is exhumed is based on the crater morphology, which is dissimilar to fresh impact craters, and the absence of visible ejecta blankets of these craters. Polygonal terrain occurs beneath the cratered terrain; grooved terrain is exposed beneath etched terrain. We admit that the number of images on which these topographic relationships are based on is rather small and it is clearly understood that more image data in combination with MOLA topography are necessary to put the proposed relationships on a statistically more sound basis.

4. Conclusions

[56] A synthesis of Mars Global Surveyor, Odyssey and Viking results leads to following conclusions: (1) gravity data [Smith et al., 1999b] most likely do not support an origin related to an impact as previously proposed by Meyer and Grolier [1977]; (2) crustal thickness is on the order of 45–60 km and significantly thinner than underneath Tharsis Montes [Smith et al., 1999b; Zuber et al., 2000]; (3) Syrtis Major lavas are ~0.5–1.0 km thick; (4) the volume of lava is ~1.6 × 10^5–3.2 × 10^5 km^3; (5) the calderas of Nili and Meroe Patera are located within a large N-S elongated central depression and are at least 1.5 km deep; (6) slopes of the flanks of Syrtis Major are ≤1° at baselines of hundreds of kilometers; (7) the very gentle slopes are similar to other Martian highland patera such as Amphitrites, Tyrhenna, and Syria Planum but are much smaller than for Martian shield volcanoes such as Olympus, Elysium, Ascraeus, and Arisia Mons; (8) the surface of the Syrtis Major Formation is significantly smoother at kilometer wavelengths than the surrounding highland plains; (9) at small wavelengths (0.6 km) Isidis Planitia appears rougher than Syrtis Major; (10) the differences in surface roughness and wrinkle ridge patterns between Isidis Planitia and Syrtis Major are likely to be related to additional sedimentation in the impact basin; (11) the wrinkle ridge pattern and topography in the southeast quadrangle of Syrtis Major might be related to loading and gravitational sector collapse; (12) Syrtis Major is heterogeneous in albedo, surface roughness, composition
and topography; (13) the smooth morphology and relatively low topography of the Isidis rim in the Syrtis Major region can be explained by initial heterogeneities and flooding with lavas, together with erosion and/or tectonic modification of the original rim; (14) our new crater counts revealed an Early Hesperian age, indicating that Syrtis Major is older than in the geologic map of Greeley and Guest [1987] and supporting ages derived by Condit [1978] and Maxwell and McGill [1988]; (15) the terrain types, their topographic relationships and the inferred stratigraphic sequence suggest large-scale erosional, possibly eolian, modification of the Syrtis Major lavas.

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