The global martian volcanic evolutionary history

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Viking mission image data revealed the total spatial extent of preserved volcanic surface on Mars. One of the dominating surface expressions is Olympus Mons and the surrounding volcanic province Tharsis. Earlier studies of the global volcanic sequence of events based on stratigraphic relationships and crater count statistics were limited to the image resolution of the Viking orbiter camera. Here, a global investigation based on high-resolution image data gathered by the High-Resolution Stereo Camera (HRSC) during the first years of Mars Express orbiting around Mars is presented. Additionally, Mars Orbiter Camera (MOC) and Thermal Emission Imaging System (THEMIS) images were used for more detailed and complementary information. The results reveal global volcanism during the Noachian period (~3.7 Ga) followed by more focused vent volcanism in three (Tharsis, Elysium, and Circum-Hellas) and later two (Tharsis and Elysium) volcanic provinces. Finally, the volcanic activity became localized to the Tharsis region (about 1.6 Ga ago), where volcanism was active until very recently (200–100 Ma). These age results were expected from radiometric dating of martian meteorites but now verified for extended geological units, mainly found in the Tharsis Montes surroundings, showing prolonged volcanism for more than 3.5 billions years. The volcanic activity on Mars appears episodic, but decaying in intensity and localizing in space. The spatial and temporal extent of martian volcanism based on crater count statistics now provides a much better database for modelling the thermodynamic evolution of Mars.

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1. Introduction

Volcanism at the surface of Mars is extensive but not uniformly distributed (Fig. 1) and includes a diversity of volcanic landforms such as central volcanoes, tholi, paterae and small domes, as well as vast volcanic plains. This is reflected in the comprehensive collection of volcanic features based on Viking images compiled by Hodges and Moore (1994). Such a diversity of landforms implies different eruption styles and probable changes in the style of volcanism with time, as well as interaction with the martian cryosphere and atmosphere during the evolution of Mars. Many volcanic constructs are associated with regional tectonic or local deformational features. Two topographically dominating and morphologically distinct volcanic provinces are the Tharsis and Elysium regions, whose morphologies are strongly analogous to basaltic volcanic landforms on Earth. Both are situated close to the equator on the dichotomy boundary between the cratered (older) highlands and the northern lowlands and are approximately 120° apart. The huge volcanoes in the Tharsis region (Olympus Mons, Ascalon Mons, Pavonis Mons, and Arsia Mons) share many characteristics with Hawaiian basaltic shield volcanoes (Carr, 1973). They are constructed from multiple lobate flows of lava (Bleacher et al., 2007; Hiesinger et al., 2007), and show complex nested coalescing summits calderas of varying age, and gentle slopes on the order of a few degrees. Their eruption style appears effusive and relatively non-explosive in nature, although there are indications of composite volcano construction (Head and Wilson, 1998; Head et al., 1998a). The main differences between the martian and terrestrial volcanoes are the greater sizes and lengths of flows of the martian examples, mainly due to higher eruption rates, the “stationary” character of the source (no plate tectonics) and the lower gravity (Basaltic Volcanism Study Project, henceforth BSVP, 1981). Plescia and Saunders (1979) summarized the chronology and classification of martian volcanic activity based on the Viking imagery data, and grouped the volcanic landforms into (1) shield volcanoes (to be of basaltic composition), (2) domes and composite cones, (3) highland paterae, and related (4) volcano-tectonic features. Many plains units like Lunae Planum and Hesperia Planum are thought to be of volcanic origin, fed by morphologically clearly-defined volcanoes or by huge fissure volcanism. Many small volcanic cone fields in the northern plains are interpreted as cinder cones (Wood, 1979), formed by lava and ice interaction (Allen, 1979; Greeley and Fagents, 2001), or as the products of phreato-eruptive eruptions (Frey et al., 1979). An overview of the temporal distribution of processes, including the volcanic activity as well as the erosional processes manifested by large outflow channels ending in the northern lowlands and sculpting large units of the volcanic flood plains is given by Neukum and Hiller (1981). Plescia and Saunders (1979) and Neukum and Hiller (1981) both gave chronological classifications of
the volcanic features on Mars. The difference between these is the applied reference crater production function. Plescia and Saunders (1979) based their reference crater production function on counts in the Lunae Planum region, although Neukum and Wise (1976) and Wise et al. (1979a) demonstrated resurfacing activity that obscured the crater size-frequency distribution of this region. Such calibration curves result in different interpretations of the stratigraphic relationships when applied on a global scale.

Based on Mars Orbiter Laser Altimeter (MOLA) topographic data, Plescia (2004) reviewed the dimensions and volumes of all the major volcanic constructs on Mars and could correct the earlier findings of the Viking era. The mapping of Spudis and Greeley (1977) and Scott and Carr (1978) indicates that as much as 60% of the surface is covered with volcanic materials. High Resolution Stereo Camera (HRSC) image data from the ESA Mars Express mission provide excellent insights into the morphology and topography of the volcanoes and their chronostratigraphic evolution. The primary focus outlined below is on the evolution of the large volcanic provinces Tharsis and Elysium, as well as other volcanic regions. An additional goal is to understand the interacting processes of erosion and deposition (related to volcanic and fluvial processes). Greeley and Spudis (1981) described the volcanic history of Mars based on the observation of superposition, cross-cutting relations, and, if available, on the number of superposed impact craters. To understand the volcanic evolution caution must be given to the fact that the amount of volcanic activity represented by the enveloping youngest layer on top of the stratigraphic sequence, and sometimes the crater size-frequency distribution reveal an earlier phase(s) by an embayed or flooded crater population. The oldest unit considered here are the highland plateau units (e.g., Wilhelms, 1974). The oldest highland-plains units are about 4.02 Ga old (Noachis Terra) that is roughly the time of formation of the largest martian basins (Werner, 2008), followed by the emplacement of the inter-crater plains at 3.9 Ga. Most of the highland unit ages range between 4.1 Ga and 3.9 Ga (Werner, 2008), roughly the end of the heavy bombardment period. At that time, the erosional scarp of the dichotomy boundary between the highlands and lowlands had most likely formed (Zuber et al., 2000). Irrespective of the cause of the dichotomy escarpment formation, subsequent resurfacing acted differently along the boundary (Werner, 2005).

The temporal overlap of extensive fluvial activity (e.g., valley networks) and volcanic episodes (e.g., highland paterae) is manifested in the phreatomagmatic interaction, e.g., at Tyrrhena Patera or at the flank base of western Elysium Mons (Wilson and Mouginis-Mark, 2003). The coincidence of the Hellas basin formation and the accumulation of highland paterae have been noted by Greeley and Spudis (1981), but this study shows that at least the later-stage patera activities were not triggered by the impact event itself. Following the interpretation of the volcanic history outlined by Greeley and Spudis (1981) based on the Viking image data, the plateau plains volcanic activity was followed by massive flood volcanism, which resurfaced very extensive areas such as Lunae or Hesperia Planum and the martian lowland areas. Massive volcanic constructs such as the Tharsis rise, notably Olympus Mons and Alba Patera, cover the dichotomy starting at least 3.8 Ga ago (the age of the aureole has been recalculated from Hiller et al., 1982) and continuing to about 3.5 Ga (Alba Patera). The presence of the aureole around Olympus Mons and the absence of such a feature around Alba Patera might indicate a changed environmental situation. The timing and existence of a martian ocean in the northern lowlands is discussed (Werner, 2005) and constrained temporally by the surface ages of its two different flank bases. Central vent volcanoes dominate both martian volcanic centers and the surrounding plains-forming flows. Fracturing (graben formation) is related to the early structural uplift of the volcanic rises, although it could have been caused by younger volcanic activity (see below). Although Greeley and Spudis (1981) found an agreement with a moon-like thermal history, a more divergent evolutionary history of Mars will be shown here. All major volcanic constructs, including paterae and tholi, have been imaged in the early phase of the ESA Mars Express mission. The ability to image simultaneously in colour and stereo provides a new opportunity to better characterize the geomorphology and chronostratigraphy of most volcanoes in the Tharsis and Elysium region and most of the highland volcanoes. Major parts of the volcanic shields and calderas were remapped on the basis of the High Resolution Stereo Camera (HRSC) images, in combination with nested images from MOC-NA. All age determination results are listed in Table 1 and a summary chart is shown in Fig. 2. The Appendix provides details on the age dating technique using crater
Table 1
Best fit relative and absolute ages of volcanic units on Mars.

<table>
<thead>
<tr>
<th>Volcanic construct</th>
<th>Area</th>
<th>Cum. number (≥1 km)</th>
<th>Ages in Ga</th>
</tr>
</thead>
<tbody>
<tr>
<td>Alba Patera Flank Regions</td>
<td>1Ar1</td>
<td>1.23e−3/3.10e−4/9.09e−5+</td>
<td>2.51/0.636+0.186+</td>
</tr>
<tr>
<td></td>
<td>2Ar1a</td>
<td>6.13e−4/2.40e−4+</td>
<td>1.26/0.493+</td>
</tr>
<tr>
<td></td>
<td>2Ar2</td>
<td>1.45e−3/3.95e−4/1.26e−4+</td>
<td>2.93/0.811+0.259+</td>
</tr>
<tr>
<td></td>
<td>3Ar1</td>
<td>5.60e−4</td>
<td>1.15</td>
</tr>
<tr>
<td>Caldera and Flank of Arisa Mons</td>
<td>cal</td>
<td>6.23e−5</td>
<td>0.128</td>
</tr>
<tr>
<td></td>
<td>1Ar1</td>
<td>1.00e−4</td>
<td>0.206</td>
</tr>
<tr>
<td></td>
<td>1Ar2</td>
<td>3.13e−3/4.10e−4+</td>
<td>3.54/0.841+</td>
</tr>
<tr>
<td></td>
<td>2Ar1</td>
<td>9.43e−4/2.11e−4+</td>
<td>1.93/0.432+</td>
</tr>
<tr>
<td></td>
<td>2Ar2</td>
<td>1.68e−4</td>
<td>0.345</td>
</tr>
<tr>
<td></td>
<td>2Ar3a</td>
<td>9.21e−5</td>
<td>0.189</td>
</tr>
<tr>
<td></td>
<td>2Ar4</td>
<td>5.09e−5</td>
<td>0.104</td>
</tr>
<tr>
<td></td>
<td>3Ar1</td>
<td>3.02e−5</td>
<td>0.62</td>
</tr>
<tr>
<td></td>
<td>4Ar1</td>
<td>5.17e−4/1.37e−4+</td>
<td>1.06/0.280+</td>
</tr>
<tr>
<td></td>
<td>5Ar1</td>
<td>9.66e−4/2.79e−4+</td>
<td>1.98/0.572+</td>
</tr>
<tr>
<td></td>
<td>5Ar2</td>
<td>4.17e−4/7.13e−5+</td>
<td>0.855/0.146+</td>
</tr>
<tr>
<td>Caldera and Flank of Pavonis Mons</td>
<td>cal 1</td>
<td>1.79e−4</td>
<td>0.367</td>
</tr>
<tr>
<td></td>
<td>cal 2</td>
<td>4.02e−5</td>
<td>0.082</td>
</tr>
<tr>
<td></td>
<td>1Ar1</td>
<td>6.32e−4/2.52e−4+8.91e−5+</td>
<td>1.30/0.516+0.183+</td>
</tr>
<tr>
<td></td>
<td>2Ar1</td>
<td>3.14e−5</td>
<td>0.064</td>
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<td></td>
<td>2Ar2</td>
<td>3.72e−4</td>
<td>0.763</td>
</tr>
<tr>
<td></td>
<td>2Ar3</td>
<td>3.36e−3/1.93e−4+</td>
<td>3.56/0.395+</td>
</tr>
<tr>
<td></td>
<td>2Ar4</td>
<td>6.03e−4</td>
<td>1.25</td>
</tr>
<tr>
<td></td>
<td>2Ar5</td>
<td>5.98e−4/1.77e−4+</td>
<td>1.23/0.362+</td>
</tr>
<tr>
<td></td>
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<td>0.212</td>
</tr>
<tr>
<td></td>
<td>3Ar3</td>
<td>4.64e−5</td>
<td>0.095</td>
</tr>
<tr>
<td></td>
<td>3Ar4</td>
<td>1.13e−4/4.33e−5+</td>
<td>0.232/0.089+</td>
</tr>
<tr>
<td>Caldera and Flank of Ascaeus Mons</td>
<td>cal 1</td>
<td>1.93e−4</td>
<td>0.396</td>
</tr>
<tr>
<td></td>
<td>cal 2</td>
<td>1.04e−4</td>
<td>0.213</td>
</tr>
<tr>
<td></td>
<td>cal 3</td>
<td>5.07e−5</td>
<td>0.104</td>
</tr>
<tr>
<td></td>
<td>cal 4</td>
<td>3.83e−4/1.13e−4</td>
<td>0.785/0.233</td>
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<td></td>
<td>cal 5</td>
<td>5.26e−5</td>
<td>0.108</td>
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<td></td>
<td>cal 6</td>
<td>5.03e−3/4.49e−5</td>
<td>3.66/0.092</td>
</tr>
<tr>
<td></td>
<td>1</td>
<td>4.87e−5</td>
<td>0.100</td>
</tr>
<tr>
<td></td>
<td>2Ar1</td>
<td>2.15e−4/7.11e−5+</td>
<td>0.44/0.145+</td>
</tr>
<tr>
<td></td>
<td>2Ar2</td>
<td>2.42e−4/8.99e−5+</td>
<td>0.496/0.184+</td>
</tr>
<tr>
<td></td>
<td>2Ar3</td>
<td>3.01e−4/1.14e−4+</td>
<td>0.617/0.233+</td>
</tr>
<tr>
<td></td>
<td>3Ar1</td>
<td>9.24e−4/8.76e−5+</td>
<td>1.90/0.179+</td>
</tr>
<tr>
<td></td>
<td>3Ar2</td>
<td>7.62e−4/1.49e−4+1.06e−4+</td>
<td>1.56/0.306+0.217+</td>
</tr>
<tr>
<td></td>
<td>4</td>
<td>5.07e−4/1.66e−5+</td>
<td>1.04/0.034+</td>
</tr>
<tr>
<td>Olympus Mons</td>
<td>The caldera vicinity</td>
<td>ar1</td>
<td>3.47e−4/1.04e−4+</td>
</tr>
<tr>
<td></td>
<td></td>
<td>ar2</td>
<td>1.13e−4</td>
</tr>
<tr>
<td></td>
<td></td>
<td>ar3</td>
<td>8.30e−5</td>
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<td></td>
<td></td>
<td>ar4</td>
<td>6.51e−5</td>
</tr>
<tr>
<td></td>
<td>Caldera</td>
<td>1.05e−4/4.95e−5</td>
<td>0.215−0.101</td>
</tr>
<tr>
<td>The eastern flanks #1</td>
<td>Plateau1</td>
<td>8.91e−5</td>
<td>0.183</td>
</tr>
<tr>
<td></td>
<td>Plateau2</td>
<td>2.52e−4/9.44e−5+</td>
<td>0.519/0.194+</td>
</tr>
<tr>
<td></td>
<td>Floor</td>
<td>8.60e−5/4.12e−5+</td>
<td>0.176/0.085+</td>
</tr>
<tr>
<td>The western flanks #2</td>
<td>Remnants</td>
<td>1.31e−2</td>
<td>3.83</td>
</tr>
<tr>
<td></td>
<td>Flanks</td>
<td>2.00e−3/5.24e−5</td>
<td>3.34−0.100</td>
</tr>
<tr>
<td>Domes northeast of the Tharsis Montes</td>
<td>Caldera</td>
<td>6.11e−3/2.44e−3+</td>
<td>3.70/3.45+</td>
</tr>
<tr>
<td></td>
<td>Shield</td>
<td>6.16e−3/3.21e−3+</td>
<td>3.70/3.54+</td>
</tr>
<tr>
<td>Uranius Patera</td>
<td>Caldera</td>
<td>1.93e−2</td>
<td>3.9</td>
</tr>
<tr>
<td>Uranius Tholus</td>
<td>Shield</td>
<td>4.71e−2/2.81e−3+</td>
<td>4.04/3.50+</td>
</tr>
<tr>
<td>Ceraunus Tholus</td>
<td>Shield</td>
<td>7.50e−3</td>
<td>3.74</td>
</tr>
<tr>
<td>Tharsis Tholus</td>
<td>Shield</td>
<td>6.63e−3/3.17e−3+</td>
<td>3.71/3.54+</td>
</tr>
<tr>
<td>Domes West of the Tharsis Montes</td>
<td>Biblis</td>
<td>5.56e−3</td>
<td>3.68</td>
</tr>
<tr>
<td></td>
<td>Calida</td>
<td>2.97e−2/7.22e−3+</td>
<td>3.92/3.73+</td>
</tr>
<tr>
<td></td>
<td>Jovis Tholus</td>
<td>No counts</td>
<td>-</td>
</tr>
<tr>
<td>Elysium Mons</td>
<td>Calida</td>
<td>2.70e−3/3.88e−4</td>
<td>3.49/1.6+</td>
</tr>
<tr>
<td></td>
<td>Northern flank</td>
<td>2.68e−3/7.86e−8</td>
<td>3.48/1.61+</td>
</tr>
<tr>
<td></td>
<td>Flank</td>
<td>6.12e−3/2.47e−3/8.00e−4</td>
<td>3.7/3.45/1.64+</td>
</tr>
</tbody>
</table>
counts, absolute ages in partially resurfaced units and errors in the relative and absolute age determination.

2. The Tharsis volcanic province surface ages

The Tharsis region is the most dominating feature of the martian topography and shows numerous volcanic constructs of different age and morphology.

2.1. Alba Patera

The northernmost volcano of the Tharsis assemblage is Alba Patera (Fig. 1), which is a gigantic volcanic shield with a base diameter of roughly 1100 km, even larger than the enormous Olympus Mons. Compared to Olympus Mons, Alba Patera is a wide but low-relief construct about 6 km in height, with flank slopes of about 1° and calderas extending to 120 km in diameter. The surrounding flank grabens (Tantalus and Alba Fossae) extend in a North–South direction. There are extensive lava flows and local dendritic valleys found at the summit region around the caldera. The volcano is divided into two parts indicating at least two formation stages; one broad lower construct (about 4–5 km high) cut by the Fossae and marginal lava aprons (Ivanov and Head, 2003), and a much smaller summit shield (about 1 km in height) which contains a caldera and is situated on top of a broad summit plateau in the shallow lower construct. Age determinations by Neukum and Hiller (1981) yielded surprisingly young ages. For comparison with the absolute ages described here, the collection of relative ages of Neukum and Hiller (1981) are recalculated on the basis of their crater retention ages $N_{\text{cum}} (D \geq 1 \text{ km})$, applying the later chronology model of Hartmann and Neukum (2001) and shown in Fig. 2.

Based on Viking and HRSC data, Werner (2005) confirmed the maximum age of about 3.5 Ga by measurements at the northern flank base of Alba Patera, an area where the dichotomy scarp cannot be seen. Measurements based on the HRSC image taken during orbit 1272 at the lower flank (north of the summit) support this oldest age deduced from the low-resolution Viking image data. At the western flank of Alba Patera, I found ages indicating at least four episodes of activity at about 3.4 Ga, 2 Ga, 800 Ma ago, and as recent as 250 Ma ago (Fig. 3). All other crater counts have been made in the upper part of the construct and indicate similar ages of about 1.1 Ga to 2 Ga. Resurfacing is possibly related to the formation of sinuous channels, which probably eroded through flowing lava as seen on other martian volcanoes, such as Hecates Tholus (Williams et al., 2005). Fluvial erosion by snow-pack melting possibly related to volcanic activity (Fassett and Head, 2007) could also explain such channel features. The youngest ages are found in the closest vicinity of the large caldera, yielding two episodes that ended as recently as between 800 Ma and 180 Ma ago. This two-stage activity is supported by the summit caldera morphol-
Fig. 2. Summary of crater frequencies $N_{\text{cum}}$ (1 km) (left scale) and model ages derived applying the cratering chronology model by Hartmann and Neukum (2001) (right scale) for most of the volcanic constructs on Mars and a few volcanic plains units (empty and filled squares). The results are plotted for each individual volcano and grouped for Highland Volcanoes, Elysium and Tharsis Region, Highlands, and Volcanic deposits/plains—Lowland Units. Most of the measurements were performed on HRSC image data (circles for caldera ages and empty squares for flank ages), other measurements on Viking MDIM 2.1 image data (filled squares, Werner et al., in preparation). Measurements on MOC-NA image data for the flank ages (triangle) constrain the small crater-size range. Three measurements by Williams et al. (2008b) from THEMIS daytime IR (upside-down triangle) are added for some highland volcanoes. The ages are listed in Table 1. For comparison, earlier measurements from Blasius (1976); Carr (1976); Carr et al. (1977); Crumpler and Aubele (1978); Masursky et al. (1977); Neukum and Wise (1976); Plescia and Saunders (1979); Wise et al. (1979b) and Neukum and Hiller (1981) based on Viking image data are plotted (stars). The latter were recalculated here correcting for the revised crater production function (Ivanov, 2001) and cratering chronology model (Hartmann and Neukum, 2001). These ages are not listed in Table 1. Horizontal lines show the epoch boundaries.

ogy, which has been interpreted by Ivanov and Head (2003) and Plescia (2004) to represent at least two major episodes of caldera formation and summit volcanic activity. Morphology and the ages suggest a complex geologic evolution of Alba Patera over most of the martian history.

2.2. The Tharsis Montes

Three large volcanoes named Ascraeus Mons, Pavonis Mons, and Arsia Mons constitute the Tharsis Montes (Fig. 1). They are centred on top of the volcanic rise as a chain trending from northeast to southwest, where Ascraeus is the northernmost and Arsia the southernmost. Previous work by Plescia and Saunders (1979) and Neukum and Hiller (1981) indicated decreasing surface ages towards the northeast. Morphology, slope steepness and caldera complexity were used as arguments to judge Arsia Mons, a broad feature with shallow slopes and large simple calderas, as the oldest, while the others were considered younger due to their steeper slopes and more complex smaller central calderas. Comparing the morphometric properties of Viking- and MOLA-derived summit elevations, large differences have been noted by Smith et al. (2001). Earlier determinations by the USGS based on Viking images (US Geological Survey, 1989) had to be reduced for all large volcanoes in the Tharsis region (Plescia, 2004).

2.2.1. Arsia Mons

The southernmost volcano of the Tharsis triplet is Arsia Mons with a summit height of about 18 km and relief height (height compared to the surrounding plains) of about 11 km, which makes it the second in terms of size to Olympus Mons. It has a single caldera, the largest on Mars, with a diameter of about 120 km (Crumpler and Aubele, 1978). The main edifice has a width of about 430 km and is composed of the central shield and two aprons at the north-eastern and the south-western flank sides, roughly following the great circle trend of the Tharsis Montes triplet. These aprons are formed by lava flows extending from alcoves on the lower flanks of the main shield and originate 5–7 km below the summit. Both aprons appear at the tip of some flank depressions, which follow a line of nine low shields (relief of about 150 m) across the caldera floor and their axis is along the same great circle trend as the Tharsis Montes (Head et al., 1998a, 1998b). While the main edifice has slope angles of about $5^\circ$, the flank apron slopes range between $1^\circ$ and $4^\circ$. At the flank base, towards the west, there is a large aureole deposit probably formed by glacial deposits (Head and Marchant, 2003). The surface age of the caldera floor is about 130 Ma (Werner et al., 2004, 2005; Neukum et al., 2004) confirming the earlier measurements of Neukum and Hiller (1981). Construct-wide ages range between 100 Ma and 200 Ma and probably represent surfaces formed during the latest stages of the summit and flank eruptions. The earlier episodes stopped at about 500 Ma, 800 Ma, and 2 Ga ago. The oldest age is about 3.54 Ga and indicates the time when the period of major edifice construction ended (Fig. 4).

2.2.2. Pavonis Mons

The middle volcano of the Tharsis Montes triplet is Pavonis Mons, which has the lowest summit altitude at about 14 km and a relief compared to the surroundings at the foot of the shield of about 10 km. The two visible caldera depressions are about 100 km in diameter and reflect the latest phases of summit ac-
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Fig. 3. The details of HRSC orbit 1272 cover the flanks of Alba Patera, on which units were counted. The scale indicates 10 km. The resulting crater size-frequency distributions are shown and the corresponding ages for the best-fit isochrons are given in Table 1.

tivity. From crater counts, the caldera floor formation ended at about 370 Ma ago for the larger one and about 80 Ma ago for the well preserved smaller caldera floor (Fig. 5a). The flanks have a slope angle of roughly 4°, but two lava aprons (similar to those seen at Arsia Mons) originate about 4 km below the summit region (slope of about 1°), aligned to the northeast-southwest regional fracture trend. Prominent lava channels and smaller alcoves carve the southern flank. Clusters of small low-shield volcanoes occupy the lava aprons and large units at the western flank bottom (towards which most of the lava flowed). These regions are associated with the youngest parts of the Pavonis Mons shield, which are reflected in many crater size-frequency distributions seen in the flanks, aprons and small shields. All ages range between about 100 Ma and 800 Ma and appear to be strongly correlated temporally with the latest stage of summit activity. Ages range between 100 Ma and 450 Ma for a series of arcuate concentric grabens that cut across the lower northwestern flank, and with lava flows barely covering the graben morphology. The main edifice was erected about 3.56 Ga ago; experiencing a strong resurfacing recorded in the crater size-frequency distribution about 1.2 Ga ago (Fig. 5b). Like Arsia Mons, there is an aureole-type deposition at the northern flank base, but no ages were determined due to the lack of visible craters in the HRSC dataset.

2.2.3. Ascraeus Mons

Ascraeus Mons is the northernmost of the volcano triplet and lies at an elevation of about 18 km. Once again, there are marginal lava aprons extending northeast and southwest at the flanks and also comparatively small aureole deposits are found at the northwestern flank segment. The flanks appear to have slope angles of about 7°. The summit caldera is complex compared to the other shield calderas and has at least six coalescing calderas. Strong tectonic features indicate reworked caldera floor morphology, which makes estimates of surface age and the caldera formation time difficult. Ages range over the entire history of the volcano, starting at about 3.6 Ga ago when the main edifice was already emplaced, to as young as ca. 100 Ma ago (Werner et al., 2004, 2005; Neukum et al., 2004). Most of the younger ages found for the caldera floors are similar to the caldera-surrounding summit area, ranging between 100 and 800 Ma, which are also found at the flanks. Some large flank units appear to be even younger (50–100 Ma; Fig. 6). A detailed investigation by Plescia (2004) reveals a complex history of flank eruption and apron formation; pronounced flow lobes cover most units. Low shield vents, alcoves and other morphologies indicating volcanic activity occur, which follow the overall northeast–southwest trend in the long axis of the triplet as already noted for Aris and Pavonis Montes.

2.2.4. Tharsis Montes

All three large shield volcanoes have a long complex volcanic history, in which the main shield formation ended at about 3.55 Ga ago (Arsia Mons: 3.54 Ga; Pavonis Mons: 3.56 Ga; Ascraeus Mons: 3.6 Ga) and which was followed by many episodes of surface modification, which covered the edifice with many layers of lava flows. Only the last period of effusive eruption is recorded by surface ages derived from crater counts. The youngest flank eruptions (which produced a huge volume) and many scattered low shield vents, indicate that the volcanoes were active until recently. The martian large shield volcanoes are probably now dormant.
2.3. Olympus mons

The largest and most prominent martian shield volcano is Olympus Mons, which has a relief of about 22 km and a basal extent of about 800 km by 600 km measured from the edges of the scarp surrounding the main edifice. The width is almost doubled if the enormous extent of the enigmatic aureole deposit is included, which is dispersed over many hundreds of kilometers.

Fig. 4. The details of HRSC orbit 957 cover the flanks of Arsia Mons, on which units were counted. The scale indicates 10 km, if not said otherwise.
Fig. 4. (continued)

Fig. 5. (a) The details of HRSC orbit 891 cover the calderas of Pavonis Mons, on which units were counted. The scale indicates 10 km. (b) The details of HRSC orbit 902 cover the flanks of Pavonis Mons, on which units were counted. The scale indicates 10 km.
north-eastward into the lowland units. The slopes are typically about 5°, and up to 30° at the scarp. Based on HRSC data, we performed extensive and detailed structural investigations of the scarp units, flanks and caldera (Neukum et al., 2004; Basilevsky et al., 2005, 2006) as well as a chronostratigraphic analysis of the entire edifice, focusing on the caldera and the western and
eastern scarp. The caldera of Olympus Mons consists of at least six coalescing depressions, which were described in detail by Mouginis-Mark (1981), and which suggest a sequence of at least six episodes of caldera collapse. Both of the smallest calderas appear very smooth in their floor morphology and were the last to form in the sequence. The other sub-caldera floors have experi-

Fig. 6. The details of HRSC orbit 68, covering the flanks of Ascraeus Mons, on which units were counted. The scale indicates 10 km.
enced strong tectonic deformation, shown by circular grabens in one of the larger depressions and wrinkle ridges that formed as compressional features spread over the centre. Additionally, most of the caldera floor has been covered by volcanic flows from effusive fissures. All caldera floor crater counts appear to cluster at around 150Ma. However, based on morphology alone, some crater counts give an erroneous inverted impression of the true sequence of events (Neukum et al., 2004). As described by Mouginis-Mark (1981), the larger calderas have been reworked by tectonic activity (graben, wrinkle ridges) and possibly volcanism. Therefore, the crater size-frequency distribution measurements of the caldera floors represent the lower age limit of the surface formation being partially resurfaced (right top). Careful mapping of the graben structures reduces the reference area, and the crater-size frequencies per area of both large calderas are shifted towards slightly older surface age (top right, before and bottom right, after area reduction).
et al. (2005) identified different slope types. One type (S2 slope) is present in most places but is occasionally modified by lava flows covering and flattening of the steep flanks. Detailed investigation constrains the sequence, while the ages derived from a set of crater size-frequency measurements yield individual fit ages of around 150 Ma, although they are statistically indistinguishable within the error limits (Fig. 7). The averaged surface age found for the Olympus Mons’ caldera is in good agreement with earlier measurements (Hartmann, 1999; Hartmann and Neukum, 2001), which were based on MOC images in small areas of the calderas of Olympus Mons as well as Arsia Mons. This indicates that the summits of these edifices have been essentially active almost up to the present.

The vicinity of the caldera is characterized by superposing lava flows, representing the latest active phases. Measuring the flank surface ages in four units around the caldera, the average surface age is oldest close to the caldera (resurfacing ended about 210 Ma ago), and ages of about 170 Ma are found further down-slope and elsewhere on the flanks. Morphologically, the coverage by lava flows is more effective further down-slope than in the close vicinity of the caldera, which is also reflected in the crater size-frequency distribution. Here, apart from a few large impact craters that are embayed by younger lava flows (one 10.5 km crater is visible in Fig. 8), the flank surface is not completely covered by flows and thus part of an older 700 Ma-aged surface has survived. This agrees with widely-observed flank eruptions on Mars and Earth. Using the observed crater size-frequency distribution, the effect of geological resurfacing is visible in the shape of the distribution (Neukum and Horn, 1976). The crater diameter (and derived rim height) at which the crater size-frequency distribution is affected, yield a thickness of an uppermost layer a few hundred meters deposited within about 500 Ma.

The summit plateau is truncated by an up to 7 km high scarp, which is present in most places but is occasionally modified by lava covering and flattening of the steep flanks. Detailed investigation of the western scarp morphology has been based on HRSC and MOC imagery (Neukum et al., 2004; Basilevsky et al., 2005, 2006). The western flank edge exposes a ridge and several smaller mesas. The ages of the mesa surface suggest that they could be remnants of the very early and ancient proto-Olympus Mons. Ages of about 3.8 Ga were determined by measurements in the aureole (Hiller et al., 1982), supporting the notion that most of the volcanic construct was emplaced very early in martian history. Prominent and thin layering is visible at the steep slopes of the scarp, indicating the volcanic origin of the plateau, and Basilevsky et al. (2005) identified different slope types. One type (S2 slope) is found only at the western flank of Olympus Mons, whilst at the upper part several chaos-like depressions occur from which channel-like grooves evolve down-slope. This morphology may relate to the influence of water (see Basilevsky et al., 2005). The ages derived from crater size-frequency measurements for the volcanic units on the flanks range between 500 Ma until very recently, indicating that the flank was blanketed by several episodes at about 500 Ma, 200 Ma, and 100 Ma. However, many of the measured crater size-frequency distributions show relatively flat distributions that indicate the steady (in terms of millions of years) supply of new lava flows, although they did not fully blanket the earliest crater record, thus putting the essential erection phase of Olympus Mons in the earliest part of martian evolutionary history (before 3.6 Ga ago).

2.4. The Tholi and Paterae on Tharsis

Eight martian volcanoes are classified as shields: Olympus Mons, Ascalco Mons, Pavonis Mons, Arsia Mons, Alba Patera, Biblis Patera, Uranius Patera, and Jovis Tholus, but a few dome-type volcanoes are also present in the Tharsis region. Northeast in the prolongation of the Tharsis Montes chain, the Uranus Group consists of Uranus Patera, Uranus Tholus, and Ceraunius Tholus. To the south and east of the Tharsis Montes lies Tharsis Tholus. These four volcanoes all have a basal diameter of between 100 and 300 km and a relief ranging from about 3 km (Uranus Patera and Uranus Tholus) to around 6 km (Ceraunius and Tharsis Tholus). Most have an asymmetric shape and a multi-stage caldera which occupies large portions of the visible construct. Many of their flanks are buried under lava flows of the surrounding plains, generated by the Tharsis Montes, so that their true dimensions remain unknown.

Uranus Patera appears to have a complex development, indicated by fan-shaped segments emanating from its complex caldera. New age determinations from low-resolution HRSC limb observation data yield a shield and caldera age of about 3.7 Ga, while a resurfacing event, which ended about 3.5 Ga ago, affected the caldera and parts of the flanks. Higher resolution HRSC images support the impression that even later resurfacing, unrelated to volcanic activity, has occurred (Fig. 9a). Uranus Tholus is a small cone-like volcano with a two-stage caldera, representing at least two active phases. New crater counts confirm that the construct had been built up at 4.04 Ga ago, while the caldera floor was em-
Fig. 9. The images on which were counted are crops of a projected limb observation during orbit 658 and details the western group of tholi and paterae sufficiently to count craters (a) Uranius Patera, (b) Uranius and Ceraunius Tholi, and (c) Tharsis Tholus. Later observations did not change the resulting ages indicating an early formation of these volcanoes.
placed before 3.9 Ga ago. A later event, ending at about 3.5 Ga ago, renewed the surface of the cone by either late stage volcanic activity or surface erosion, which cannot be evaluated at the image resolution given (Fig. 9b). Ceraunius Tholus is striking in the prominent sets of radial troughs at its flank, partly leading into the elliptical crater Rahe that was filled with lava piles through the troughs. A two-stage caldera indicates episodic activity, which is not revealed by crater counting. The volcanic construct was emplaced about 3.75 Ga ago (Fig. 9b). Tharsis Tholus has a complex morphology and is unique among martian volcanoes because of the slumping blocks that segment its flanks. The volcano is an obstacle in the surrounding lava plains, formed by lava flows from Ascraeus Mons. The age determined by crater counts indicates that the visible part of the edifice was emplaced no later than about 3.71 Ga ago (Fig. 9c).

West of the Tharsis Montes is another group of volcanoes: Biblis Patera, Ulysses Patera, and Jovis Tholus (Fig. 1). All are completely surrounded by younger lavas originating from Tharsis Montes. The small volcanoes west of the Tharsis Montes became inactive before the latest stage of the Tharsis Montes activity. Only parts of the edifices are exposed at the surface, which is reflected by their low reliefs of 1 to 3 km above the surrounding plains. Biblis Patera's visible tip extends roughly 130 km by 180 km. Its asymmetrical exposure and caldera floor, lying 1–1.5 km below the surface of the surrounding plains, may provide a clue to the original extent of the edifice. Age determination for the caldera floor indicates that the remaining uppermost region formed before 3.68 Ga ago (Fig. 10). Ulysses Patera's morphology suggests that most of the edifice was buried under subsequent lava flows formed by Arsia Mons. Two large impact craters are visible on the flanks of the relatively small edifice. Crater counts on the flanks and the caldera yield an end of the edifice construction period at about 3.73 Ga ago. Nevertheless, the presence of the two large craters implies that the edifice was emplaced even earlier (about 3.9 Ga ago (Fig. 10)). Jovis Tholus, located further northeast, is an obstacle (embayed by lava flows of Arsia Mons) standing only 1 km above the surrounding plains. The caldera occupies most of the remaining cone, indicating large amounts of lava embaying the edifice but no crater counts have been made for this volcano.

The measurements presented here indicate a more homogeneous picture of the evolution of the small Tharsis volcanoes than the data collected by Neukum and Hiller (1981) due to the varying counting strategies, differing operators and variable image quality. In this paper, only a single operator performed all the measurements at a uniform image resolution. The overall impression is that the active period of the Tharsis tholi and paterae stopped early in martian history (before 3.7 Ga ago) and about 150 million years later the large volcanoes (Olympus and Tharsis Montes) completed their erection phase. Only a few of the small volcanoes show a distinctive resurfacing age at about 3.5 Ga, which is not clearly related to volcanic activity.

3. The Elysium volcanic province ages

The Elysium region is the second-largest volcanic province on Mars, situated in the northern lowlands. The volcanic province consists of three volcanoes: Elysium Mons, Hecates Tholus, and Albor Tholus. A fourth volcano, Apollinaris Patera, is located at the dichotomy boundary further south and is surrounded by parts of the Medusae Fossae formation (Fig. 1). The Elysium volcanoes and Apollinaris Patera are classified as dome and composite volcanoes respectively, and considered to be constructed from lavas that are more viscous than ordinary basalts. The domes are formed by multiple lava flows and the composite volcanoes are possibly constructed by interbedded lava and pyroclastic material. Although not confirmed, the Medusae Fossae Formation is probably made up of pyroclastic deposits, spread widely at the dichotomy boundary but with no clearly identified source.

Malin (1977) described the Elysium region in detail based on Mariner 9 data. The three Elysium volcanoes are situated on top of a broad (about 1600 km wide) regional high similar to the Tharsis rise. The structural situation has been compared to Alba Patera (Head et al., 1998b) since the vent volcanoes are superimposed on the plateau in an off-centre fashion. While Elysium Mons occupies part of the plateau, Hecates Tholus lies at the north-eastern base and Albor Tholus at the south-eastern margin. Both show some embayment contacts at their bases, which hide their complete extent.

3.1. Elysium Mons

The largest of these three volcanoes is Elysium Mons, with a relief of about 14 km and flank slopes ranging between 1° and 10°. The simple summit caldera has a diameter of about 14 km and a shallow appearance. Radial and concentric troughs dominate the western (and partly eastern) flank of the regional high, but may not be related to Elysium Mons. Along the western flanks, lava
flows associated with Elysium Mons extend several hundred kilometers into the northern lowland plains (Utopia Planitia). Fluvial features at the base of the western flank along the Elysium rise could have formed by tectonically-driven release of ground water or phreatomagmatic interactions. At the base of the south-eastern flank, traces of tectonic grabens extend further in a radial fashion (Cerberus Fossae). They may have originated during the formation of the Elysium rise and are possibly still active, cutting through the youngest region of Mars, Elysium Planitia and the Cerberus plains, (Berman and Hartmann, 2002; Werner et al., 2003a, 2003b). This area is argued to be the youngest volcanic plains unit of Mars.

The ages based on crater counts from HRSC images indicate the final activity of the emplacement of the main edifice at the latest 3.5 Ga ago, while the frequency of a few large craters yields an even older age (3.7 Ga). Both caldera and flank crater size-frequency distributions indicate a resurfacing event ending about 1.6 Ga ago (Fig. 11). This is unrelated to the Elysium Mons vent activity, but is possibly an aeolian overprint. Most measurements based on Viking images confirm that the main construction phase had ended about 3.7 Ga ago and only the large flank eruption, flowing northwest and covering Utopia Planitia had formed over a period of 400 Ma and continued to as recently as 3.1 Ga ago (Werner, 2005).

3.2. Albor Tholus and Hecates Tholus

The southernmost dome of the three Elysium volcanoes is Albor Tholus. The caldera diameter is about 35 km with a depth of 4 km, which is enormous with respect to its basal extent of about 150 km and relief of about 5.5 km. While the flanks appear convex, with slope angles of about 5°, the caldera-wall slopes range between about 20° and up to 35°. The caldera morphology indicates a smaller, younger caldera collapse. Ages found for the caldera floors indicate that the summit activity had ended about 500 Ma ago, with an earlier episode ending 1.6 Ga ago (Werner et al., 2004, 2005; Neukum et al., 2004). The edifice itself was built

**Fig. 10.** Similar to the western group of tholi and paterae, Biblis and Ulysses Paterae were covered during limb observations in orbit 629. On the projected image, sufficiently high details to count craters were found.
before 3.4 Ga and likely the caldera floor ages represent later aeolian coverage.

The northernmost volcano is Hecates Tholus, located at the edge of the Elysium rise and connected to the lowlands. Numerous radial rills emanate from the summit, originally interpreted as fluvial in origin (Gulick and Baker, 1990), but more recently viewed as eroding through lava flows (Williams et al., 2005). With a relief of about 7 km, the roughly 180 km wide edifice exposes a small (13 km in diameter) and shallow (less than 500 m) multi-staged caldera. The crater counts indicate a history of summit activity over the last billion years, while the flank age of at least 3.5 Ga suggests that the construct had been already emplaced at that early stage. Mouginis-Mark et al. (1982) studied Hecates Tholus extensively and suggested that the summit is covered with pyroclastic deposits post-dating the flank formation. The flanks appear convex, with slopes varying from 6° at the bottom to 3° at the summit. Detailed age determination for the northern flank of Hecates Tholus was initiated to understand the timing of the summit caldera activity and a possible side-caldera (Hauber et al., 2005), not discussed in earlier investigations. Images and topographic data from the HRSC revealed previously unknown traces of an explosive eruption at 30.8° N and 14.98° E, on the northwestern flank of Hecates Tholus. The northwestern flank has been mapped by us and studied in detail in terms of morphology and crater size-frequency distributions. Additionally, MOC-NA image data have been used. We found that both caldera and flanks have a similar-aged overprint. While the shape of the caldera indicates an active phase about 1 Ga ago, we find similarly aged deposits on the upper flank segment. Later caldera activities are not reflected in the surface ages of the summit vicinity, which indicates that they were less massive than the one occurring about 1 Ga ago. The eruption at the flank bottom created a large, 10-km-diameter caldera about 350 Ma ago. In the vicinity of the amphitheatre-like depression erupted deposits are found, which show this age.

Glacial deposits partly fill the caldera and a younger adjacent depression.

3.3. Apollinaris Patera

This volcano is situated at the dichotomy boundary adjacent to the impact crater Gusev and is surrounded by the Medusae Fossae Formation. It extends over roughly 190 km and has a two-stage caldera of about 80 km in diameter. A small scarp facing towards the northern lowlands characterizes the northern rim, resembling a small version of the Olympus Mons scarp. To the south, a lava fan has evolved, making up the youngest unit of the entire construct. Ages derived from crater counts indicate that the last activity (at the fan and caldera) ended at about 3.71 Ga ago, while the entire volcano had been constructed by 3.81 Ga ago (Fig. 12) with a resurfacing event at about 3.74 Ga ago. The small caldera floor had formed about 3.6 Ga ago. A similar sequence of events for Apollinaris Patera is described by Robinson et al. (1993), based on stratigraphic relations only. Therefore, Apollinaris Patera exemplifies the applicability of crater counting methods as a valuable procedure.

Even if the structure of these four volcanoes suggests assignment to one group, their evolutionary history, especially of Apollinaris Patera, is diverse. A few small Noachian-aged volcanic domes (e.g., Zephyria Tholus) are located in the highland vicinity of Apollinaris Patera (Stewart and Head, 2001), which are consistent with a strato-volcano origin in which the edifice formed by mixed explosive and effusive eruptions. These observations make Apollinaris Patera unique when compared to other highland paterae.

4. The Medusae Fossae Formation

Along the equator between the Tharsis and Elysium volcanic centers in the Elysium–Amazonis Planitiae region, the highland-
lowland boundary is superimposed by massive deposits. These deposits define an extensive unit of somewhat enigmatic origin, named the Medusae Fossae Formation (Scott and Tanaka, 1982, 1986). In general, the Formation appears as a smooth and gently undulating surface, but is partly wind-sculpted into ridges and grooves. MOLA-based topographic data indicate a thickness of the deposit of up to 3 km. It is commonly agreed that the materials forming Medusae Fossae were deposited by pyroclastic flows or similar volcanic air-fall materials (Greeley and Guest, 1987), but they have also been proposed to be remnant of ancient polar-layered deposits of a polar cap (Schultz and Lutz, 1988).

Crater counts were performed westward of Arsia Mons close to the equator, at the edge of the highland–lowland boundary and close to Apollinaris Patera. The former area between 140° W and 150° W was imaged by the HRSC during orbit 895 and orbit 917 (Fig. 13). Here, the highland–lowland boundary plateau is partly covered by lava flows and partly dissected by valleys, which were most likely carved by fluvial activity. The remains of water-bearing inner channels are visible in the center of the valleys and at the bottom of the massif. Superposition of the lobate-fronted pyroclastic flows indicates that the water erosion ended before deposition. Crater size-frequency distribution measurements for the lava-covered plateau section and parts of the pyroclastic-flow units that form Medusae Fossae, resulted in ages of about 3.1 Ga for the last lava coverage of the plateau and about 1.6 Ga for parts of the Medusae Fossae Formation. The latter age is also found as a surface age for Medusae Fossae Formation units in the Apollinaris Patera vicinity (Fig. 13). The age correlation is found also for the aeolian features at Elysium Mons and Albor Tholus. A single global formation of the enigmatic Medusae Fossae Formation around 1.6 Ga ago is most likely.

5. Highland Paterae

This category of central vent volcanoes has been defined by Plessia and Saunders (1979). Hadriaca and Tyrhena Patera, in ad-
The region, named Malea Planum, south of the Hellas impact basin, has a morphology typical of volcanic plains (e.g., Hesperia Planum) and displays the two well-defined calderas of Amphitrites and Peneus Paterae. They have low reliefs and very shallow flank slopes of about 1°. Other features in the plains unit are caldera suspects (Peterson, 1977), now confirmed and named Malea and Pityusa Paterae. HRSC observation confirmed the Malea Patera, for which crater counts yield an age of about 3.75 Ga and a resurfacing event ending around 3.6 Ga ago (Fig. 15). This event was probably unrelated to volcanic activity. Later crater counts on THEMIS daytime IR performed by Williams et al. (2008b) yield ages of about 3.8 Ga for the surfaces of Peneus, Malea, and Pityusa Paterae, and subsequent resurfacing.

Syrtis Major Planum, a volcanic plain unit, was identified as a large volcanic region west of the Isidis impact basin (Schaber, 1982). Syrtis Major occupies an area of about 1100 km in diameter and is characterized by two calderas, Meroe and Nili Paterae (Hiesinger and Head, 2004). They are located almost at the centre of the very low shield inside an elliptical depression, which was not interpreted by Hiesinger and Head (2004), but appears to resemble an earlier-stage caldera. The plateau is characterized by wrinkle ridges similar to other volcanic plains (e.g., Hesperia Planum). Ages determined for the approximately 70 km diameter caldera of Meroe Patera yield an age of about 3.75 Ga (Fig. 16), which probably indicates the end of the main volcanic activity. The crater size–frequency distribution measured for the caldera in-
Fig. 15. The details of HRSC orbit 30 cover the caldera of Amphitrites Patera, on which craters were counted. The scale indicates 10 km.

Fig. 16. The resulting crater size-frequency distribution for the measurement in the caldera of Meroe Patera (Syrtis Major), as observed in HRSC orbit 1076, is given here.

dicates a later resurfacing ending about 2.3 Ga ago, which here is interpreted as non-volcanic. Similar ages have been found by Hiesinger and Head (2004). No detailed crater counts were performed for Nili Patera.

6. Volcanic plains

Four types of martian volcanoes and their prevailing occurrences in the cratered terrain hemisphere (generally the southern hemisphere) have been previously described in the literature. Paterae, which are large low-profile volcanic structures, appear to be either older shield volcanoes or a unique type of volcano. 'Plains' volcanic units represent low-volume eruptions that formed cones, low shields, and other small-scale structures. Flood volcanic units are produced by high-volume eruptions, post-dating the older and more degraded plateau plains, and occur mostly as basin-filling materials. Plateau plains, the martian inter-crater plains, contain many wrinkle ridges and floor-fractured craters. It has been suggested that volcanic processes as well as erosional processes have been important in obliterating small martian craters (Greeley and Spudis, 1978). Further, volcanic products may constitute a significant fraction (up to 44%) of the surface rocks in the cratered terrain (Greeley and Spudis, 1978). One of the typical examples of volcanic plains is Hesperia Planum. Typical wrinkle ridges are interpreted as resembling the morphology of similar features on the lunar maria. Many ridges follow an irregular pattern, but some appear mostly circular and cover what are likely crater rims of a cratered plains unit that has been subsequently covered. Crater size-frequency measurements performed on Viking and THEMIS image data suggest that the plains were emplaced before 3.71 Ga ago, but experienced a resurfacing event about 3.12 Ga ago ($N_{\text{cum}}(\leq 1 \text{ km}) = 1.60e-3$, Werner, 2005). This emplacement age is valid for the entire Hesperia Planum unit. Similarly, Malea Planum is dated to about 3.75 Ga (Williams et al., 2008b). Another plains unit, a small region of Lunae Planum, yields an age of 3.5 Ga (Werner, 2005). The oldest unit on Mars, the highland plateau Noachis Terra, formed before 4 Ga (Werner, 2005).

7. The volcanic evolutionary history—Summary

The volcanic evolution of most individual volcanic constructs on the martian globe is described above. Detailed geological studies
have been accompanied by age dating. Those results are included in the discussion presented here. The crater count results and derived relative and absolute ages of this study are listed in Table 1, and these along with published results (Basilevsky et al., 2005, 2006; Hauber et al., 2005; Neukum et al., 2004; Werner et al., 2003a, 2003b, 2004, 2005; Werner, 2005; Williams et al., 2007, 2008a) are plotted in Fig. 2. All resulting crater counts gathered in this work are presented together with measurements based on Viking image data (Fig. 2), which were assembled by Neukum and Hiller (1981), who evaluated measurements published by Blasius (1976); Carr (1976); Carr et al. (1977); Crumpler and Aubele (1978); Masursky et al. (1977); Neukum and Wise (1976); Plescia and Saunders (1979); Wise et al. (1979b) and their own.

The novelty of the global data set presented and interpreted here is that counts have been performed by a single observer, they are based on a single set of high-resolution image data, and a single applied crater production function and chronology model. This guarantees a very coherent set of data and allows for a better comparison of the results and subsequent comparison with earlier measurements. Based on the results shown in Fig. 2, the following volcanic evolution for Mars is derived.

All volcanic constructs were formed and built up to their present size early in martian history (before about 3.6 Ga ago). This implies that most of the volcanism had started about 4 Ga ago or even earlier. In that time period, the highland units had already been emplaced and inter-crater plains were formed. Possibly even the basement of the lowlands had formed about that time (before 3.8 Ga ago). The small tholi and paterae in the Tharsis region are relics of such an early volcanic period. In their case, the vent production of magma stopped about 4 Ga ago and shows younger activity until about 3.7 Ga ago in only a few cases. Few of the tholi and paterae show later resurfacing, which is not clearly related to volcanic processes. Global volcanic activity during the early martian history is manifested in many small shields or domes found for example in the vicinity of Apollinaris Patera or in the Coraics Fossae, where we obtained ages of about 3.9 Ga for the latter (Grott et al., 2005). A similar decrease in activity is found for the highland volcanoes: Meroe Patera (3.73 Ga; Syrtis Major), Apollinaris Patera (3.71 Ga), and Amphitrites Patera (3.74 Ga), with no obvious volcanic activity occurring later than 3.6 Ga ago. Subsequent resurfacing is observed and could be related to a blanketing deposition of unknown origin which is seen both inside and outside the calderas. A more varied timing of events is observed for Hadriaca and Tyrrhena Paterae, situated east of Hellas in Hesperia Planum, a volcanic plain that formed about 3.71 Ga ago. The evolutionary history of both volcanoes appears synchronized in processes and timing. The very shallow broad shields, composed of material from more explosive volcanic activity, were already constructed by about 3.9 Ga ago. Subsequent activity resurfaced the vicinity by possible ash deposition and by later carving channels into the unstable ash deposits until about 3.7 Ga ago. The eruption style later changed to a more effusive one as observed at their crest regions (about 3.5 to 3.3 Ga ago), and as indicated by the large caldera-flank channel on Tyrrhena Patera. Inside the Hadriaca caldera and at Tyrrhena Patera, the youngest construct-wide resurfacing event ended about 1.6 Ga ago. Ages reported by Williams et al. (2008b) for Malea, Peneus, and Pituya Paterae indicate that they formed before 3.75 Ga ago. The Elysium volcanoes formed before 3.7 Ga ago and reached their final size about 3.5 Ga ago. Extensive volcanic activity is observed at the western flanks of the Elysium rise during a period between 3.4 and 3.3 Ga ago. Enormous volcanic flows expand into Utopia Planitia, while flank failures support the formation of channels by the release of water. The surrounding plains to the northeast (Arcadia Planitia) indicate subsequent volcanic flooding about 2.6 Ga ago, but the source remains unclear. The caldera and upper flanks of Elysium Mons formed about 3.5 Ga ago, showing a resurfacing about 1.6 Ga that cannot be related to volcanic activity because both ages are found inside and outside the caldera for both Hadriaca and Tyrrhena Paterae. Measurements especially at Hecates Tholus but also at Albor Tholus indicate that volcanic activity occurred later than previously thought. Caldera and flank ages show subsequent resurfacing over the past 2 Ga, while at Hecates Tholus this happened over the past 1 Ga until about 100 Ma ago. Elysium Planitia, a region southeast of the Elysium rise, is one of the youngest plains on Mars (formed between 10 and 30 Ma ago). The ages measured in this region have been interpreted as a result of the discovery of crater Zunil and its numerous secondary-crater strewn field (McEwen et al., 2005). As discussed in Werner et al. (2009), the misinterpretation in age is less than a factor of two if secondary craters were included in the measurements unwittingly.

For the other large Tharsis volcanoes, crater size-frequency distributions indicate that they reached their final size about 3.55 Ga ago, with later more or less intensive resurfacing of the flanks. At all shield flanks, episodes of volcanic eruptions (flows) are observed, which have surface ages of between 500 Ma and 100 Ma. The youngest ages determined by the crater size-frequency measurements are about 2 Ma (Olympus Mons escarpment), suggesting that the volcanoes are potentially still active (Neukum et al., 2004). For Alba Patera, the most recent flank resurfacing occurred about 200 Ma ago, possibly as the result of volcanic activity. Caldera floor ages of the Tharsis Montes and Olympus Mons show that the latest vent activity happened between 200 Ma and 100 Ma ago. At the foot of Pavonis Mons, small shields are observed that formed about 300 Ma to 100 Ma ago. This latest extensive volcanic activity, observed in both large volcanic provinces (Tharsis and Elysium), correlates well with crystallization ages found for basaltic martian meteorites (Shergottites) and indicates that the applied chronology model accurately reflects the surface ages. However, recently Bouvier et al. (2005, 2008) argued that the young Shergottite ages represent hydrothermal alteration and the crystallization age is 4.1 Ga.

This study is the first to constrain the age for the formation of the Medusae Fossae at about 1.6 Ga. This correlates with late-stage resurfacing of some highland volcanoes (Hadriaca and Tyrrhena Paterae) and at least some ages found at the flank of the Tharsis Montes. The same age is observed for some volcanic surfaces where no clear source can be identified (Elysium Mons, Albor Tholus, Meroe Patera). If the Medusae Fossae deposits are interpreted as pyroclastic ash that is widespread over the planet and accumulated at the dichotomy boundary, it is possible that this material could have been deposited as a thin layer in many places, for example, at the volcanoes. This possible final stage of global volcanic activity is perhaps supported by the age of another group of martian meteorites (Nakhitites). Crater size-frequency measurements confirm that most edifices were constructed over billions of years and are characterized by episodic phases of activity that continued in both large volcanic regions almost to the present. A number of caldera floor and flank ages cluster around 150 Ma, indicating a relatively recent peak activity period that practically coincide with radiometrically measured ages of Shergottites. The relation to the second group of martian meteorites, the Nakhitites, remains more speculative. Most of the smaller volcanoes in the Tharsis region were active in early martian geological history, similar to most of the highland volcanoes. The long activity of martian volcanoes correspondingly implies a long lifetime of the “feeding” source especially in the Tharsis region that indicates a long and stable dynamic regime in the planet’s interior.

8. Implications

The volcanic and related tectonic activity sheds light on the internal dynamics and thermal evolution of planets. Essential time-
markers are set by understanding the evolutionary history of martian volcanic constructs. After providing a time-frame for surface processes, the timing for the thermodynamical evolution of Mars can now be assessed. The apparent absence of plate tectonics on Mars and the presence of magnetized crustal materials, requiring the presence of a strong ancient dynamo field (Acuña et al., 1999), makes Mars an interesting planet to compare with Earth or other inner Solar System bodies. Many aspects of the scenario of the early and subsequent evolution, including internal dynamics of Mars, have been discussed earlier (most recently by Spohn et al., 2001; Solomon et al., 2005; Nimmo and Tanaka, 2005). Usually, the timing is based on educated guesses, while the results presented here essentially push the time-frame forward. In the early martian history, it is evident that volcanism occurred planet-wide and all volcanic regions were emplaced before about 3.5 Ga ago. It is also demonstrated that the Elysium and Tharsis regions experienced volcanic activity until very recently (i.e., 200 to 100 Ma ago, even less in a few places), suggesting than the most recent activities (over the last 500 Ma of martian history) are more widespread than previously thought.

Hand in hand with the understanding of the time frame of the decay of the martian magnetic field (Werner, 2008), the thermal evolution and the internal dynamics are more constrained. While large-scale convection in the mantle is responsible for the continuous reshaping of Earth’s surface through plate tectonics and volcanism, plate tectonics is currently not observed on Mars. The absence of plate tectonics and lower martian gravity could be a reason for the larger size of the martian volcanoes (BVSP, 1981).

On Mars, volcanism is extensive but not uniformly distributed, and includes a diversity of volcanic landforms. The Tharsis and the Elysium regions are two volcanic provinces that are topographically dominating (Fig. 1), situated close to the equator on the dichotomy boundary between the cratered (older) highlands and the northern lowlands (about 120° apart). A possible correlation of young volcanism and areoid (martian geoid) highs could suggest mantle upwelling (Steinberger et al., 2009). The regions where upwelling might be indicated in the areoid are characterized by volcanoes (Olympus Mons, Ascraeus Mons, Pavonis Mons and Arsia Mons) whose morphologies are strongly analogous to volcanic landforms on Earth.

In order to explain the internal dynamics of planets, especially that of Mars, the following scenario, summarized by Spohn et al. (2001) is required: After accretion, a planet differentiates into a core, mantle, crust and atmosphere/hydrosphere of unknown conditions. Several accretion and differentiation models have been postulated and usually incorporate time scales in the order of 20 to 50 Ma. Mars’s mass and moment of inertia factor today supports a differentiated structure. The subsequent thermal evolution or cooling history is transferring heat by conduction and by thermal convection to remove the heat generated by radioactive nuclides and to cool the interior. Large-scale convection or cooling history is transferring heat by conduction and by thermal convection to remove the heat generated by radioactive nuclides and to cool the interior. Large-scale convection has been considered to form a chalcedony crust as derived from Mars Global Surveyor gravity and topography data (Zuber et al., 2000). Numerical models show, simultaneously to a cooling core, that mantle flow is dominated by widely distributed down-welling. Broad local upwelling flows are commonly found in internally heated convection models (Spohn et al., 2001 and references therein). In the case of Mars, a hotter (or superheated) core probably allowed for a strong dynamo in the early history of Mars. A spinel-perovskite phase transition, even though Mars’s temperature and pressure schemes are not well-defined, might have amplified large-scale and localization of upwelling flows. A cooling core reduces the plume activity with time. Theses scenarios are discussed by Spohn et al. (2001) who concluded that models that attempt to explain the martian evolution are still limited by many aspects such as dynamo theory, rheology, and mantle-lithosphere interaction. Recent volcanism might no longer be driven by up-welling mantle convection (plumes) but due to crustal thickness growth and buoyancy (e.g., Schumacher and Breuer, 2007). Both ideas compete but no final proof for one or the other case can be given. In this sense, Earth appears unique among the terrestrial planets in possessing plate tectonics. Possibly, its mantle convection regime produces convective stresses to generate failure in the rigid surface boundary layer. Other planets (e.g., Venus) appear to be in a stagnant-lid regime; a regime which is characterized by the formation of a nearly immobile lid on top of a convective mantle that occurred due to large viscosity of the upper thermal boundary layer.

For Mars, the following sequence of events is suggested: Major (voluminous) volcanic activity occurred globally. The cessation of the magnetic dynamo (before 3.9 Ga ago) was followed by possible plume activity in all volcanic centers (Tharsis, Elysium, circum-Hellas and Syrtis Major region). The last observed global volcanic activity ended around 3.6 Ga ago. Plains-forming activity ended even earlier (3.7 Ga). Locally, more recent activity is observed in Elysium, Tharsis and possibly at Hadriaca and Tyrrhena Paterae. Models suggest that the pattern of mantle convection changed to a dominantly broad weak upwelling after the core cooled and the dynamo ceased as a consequence of too-low heat flux across the core–mantle boundary. Three broad areas of possible upwelling can be identified in the areoid (Steinberger et al., 2009). Volcanism and magmatism could continue as on Venus, localized in places where thick insulating crust formed (Schumacher and Breuer, 2007).

A comparison of dynamical patterns reflected on the planet’s surface and its gravity anomalies and geoid patterns allow for a better understanding of the thermal evolution of planets. Work presented here strongly support that Mars is a planet which takes its place between the two end-members of terrestrial planetary evolution: While Mercury and the Moon appear to record the early history of the Solar System, and show no signs of water relevant to their evolution, Mars at least shows a very diverse surface history. It is a planet whose surface features show the total record of both internal and external processes from the beginning of its existence 4.5 Ga ago until now. Similarly, Barlow (1988) concluded from crustal size-frequency distribution analysis on the basis of Viking image data, that Mars is the only place in the solar system to display surfaces from throughout the planet’s history. The role of water in Mars’s early history is unclear, but evidence for surface action of water is prominent (Carr, 1996). Venus and Earth are two end-members that show very young surfaces too. The major difference is the presence of water. While Venus is presumably dry, the hydrosphere on Earth might be key to plate tectonics. The validity of theories concerning (deep) plume activity and plate tectonics versus stagnant-lid and crustal thickening could gain additional support from the comparison between planets. Ultimately, these results will contribute to the knowledge of similarities and differences between planets, and thus improve our understanding on how each one of them evolves thermally in its interior and geologically on its surface.

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Appendix A. The age dating technique using crater counts

The theoretical concept and mathematical background of age dating techniques was developed in the 1960s and 1970s and summarized by the Crater Analysis Techniques Working Group (Arvidson et al., 1979). Cumulative crater size-frequency distributions are used to derive the relation between crater size-frequency distribution and relative ages on planetary surfaces. To transfer this relation from the Moon, where the crater frequency–age relation is best established, a projectile population represented by its mean-velocity distribution is used. Averaging the velocity distribution by means of the projectile mass distribution \( n(m, t) \) in a mass interval \((m, m + dm)\) results in a crater size-frequency distribution \( n(D, t) \) in the crater diameter interval \([D, D + dD]\) for a specific exposure time \( t \). The differential cratering rate \( \phi(D, t) \) is the number of craters for a specific diameter \( D \) per unit area per time at a given time \( t \). The crater size-frequency distribution of a surface unit exposed to bombardment over a time \( t \) has a relative age \((t > 0)\) described by the differential crater size-frequency distribution \( n(D, t) \):

\[
n(D, t) = \int_0^t \phi(D, t') dt'.
\]

(A.1)

Crater size-frequencies distributions integrated over crater diameters leads to the cumulative crater size-frequency distribution, i.e., craters equal to or larger than a given diameter \( D \) formed during time \( t \) on a planetary surface in its continuous approximation:

\[
N_{\text{cum}}(D, t) = \int_0^t \int_D^\infty \phi(D', t') dD' dt' = \int_D^\infty n(D', t) dD'.
\]

(A.2)

In reality, differential and cumulative frequencies are derived from discrete numbers. Following suggestions by the Crater Analysis Techniques Working Group (Arvidson et al., 1979), to display measured crater size-frequency distributions comparatively, data presented here are shown in plots with axes in double logarithmic scale with a base of 10 and equal decade length, consistent units on both axes, crater diameters in kilometer, all frequencies given per square kilometer, and 1σ-standard deviation of each measurement point (given by \( \sigma \approx \pm 1^{1/2} \), where \( n \) is the number of craters of a given crater diameter assuming a Poisson-distribution). For a better comparison, the frequency distributions are standard-binned by diameter intervals and normalised for the differential distribution. The cumulative distribution is the sum of discrete numbers of an interval \([D_a, D_b]\) are plotted as \( D_i = D_a \).

Crater size-frequency distributions measured on geological units of different size (with area \( A \)) are scaled for each bin \([D_a, D_b]\), here the cumulative crater size-frequency distribution \( N \):

\[
N_i = \sum_{k=1}^{i} \frac{n_k}{A_k}
\]

(A.3)

and plotted versus \( \log(D_b) \). It allows comparing crater populations measured on different sized surface areas and different image resolutions. The level of uncertainty of the scaled cumulative number \( N \) per bin is given by:

\[
\pm \sigma N = \log \left[ \frac{N \pm N^{1/2}}{A} \right],
\]

(A.4)

for each bin.

A1. Relative and absolute ages

The transfer of relative into absolute ages is based on the assumptions that the so-called crater production functions is known, the flux of projectiles onto the surfaces isotropic, and any target influence negligible, then the crater frequencies measured on planetary surfaces exposed for the same time with respect to diameter are the same. Crater frequencies representing different aged surfaces (at \( t_i \) and \( t_j \)) can be compared by their ratios: \( N_{\text{cum}}(D, t_i)/N_{\text{cum}}(D, t_j) = F(t_i)/F(t_j) = C \). Cumulative frequencies differ by a factor \( c_{ij} \), which is related to their age difference, and represent the relative age. For a better comparison, relative ages based on \( N \propto F(t) \) are given for a fixed diameter \( D \) (e.g., 1 km, 4 km, 10 km or 20 km), so-called crater-retention ages (Hartmann, 1966; Neukum and Wise, 1976; Neukum and Hfiler, 1981; Neukum, 1983). Based on the crater-retention ages and the relation between \( N_{\text{cum}} = G(D) \cdot F(t) \), absolute surface ages are derived applying crater chronology models which are well known for the Moon and can be transferred to other planetary bodies (Ivanov, 2001). The absolute ages or cratering model ages used here are calculated from crater-retention ages for a reference diameter \( D = 1 \) km.

Appendix B. Absolute ages in resurfaced units

Three phenomena, saturation, contamination by secondary cratering and geological resurfacing, cause characteristic deviations from the crater production function. In densely cratered units (old), crater destruction and production eventually reach a steady state, starting at small sizes. Such an equilibrium distribution exhibits a characteristic cumulative slope of “minus-two.” For surfaces with complex geological history the crater size-frequency data reveal resurfacing events as kinks and require multiple curve fits (e.g., Neukum and Hfiler, 1981; Werner, 2005). The determination of absolute ages on Mars is based on the crater production function (Ivanov, 2001, originally in Neukum, 1983) and the applied chronology model (Hartmann and Neukum, 2001).

Contamination by obvious secondary cratering (clustered, etc.) results in steeper crater size-frequency distributions than the applied crater production function predicts.

Usually, resurfacing erases the cratering record starting at the smaller size range or resets the cratering record totally, when the surface is completely renewed. Thus, it is dependent on the magnitude of the event. When measuring the crater size-frequency distribution of such a unit, all craters irrespective of their modification state are counted, any distinct drop in the distribution reveals a resurfacing event. In cases of ongoing resurfacing (e.g., on steep slopes of martian surface structures) or saturation, the crater counts cannot be linked to an age.

For units where resurfacing has been observed, the determination of ages has to be limited to a certain diameter range (below or above a diameter \( d^* \), where the resurfacing has influenced the crater size-frequency distribution). The oldest age is simply derived by fitting the production function to the large-diameter range. For the resurfacing event, if no correction has been applied, the age is slightly overestimated (depending on the magnitude of the resurfacing event) due to the cumulative character of the crater size-frequency distribution. As a first attempt, a cut-off limit (Neukum and Hfiler, 1981) applied to the large size range reduces the number of craters in the smaller size range, which is used to estimate the resurfacing event. This approach underestimates the age (depending on the surface age difference). A more precise treatment is developed by Werner (2005).

The general description of the cumulative crater size-frequency distribution \( N_{\text{cum}}(D, t) \) is given by the cumulative number of craters per area for a given diameter \( D \) at a certain time \( t \):

\[
N_{\text{cum}}(D, t) = \int_D^\infty n(D', t) dD' = \int_0^t \int_D^\infty \phi(D', t') dD' dt'.
\]
\[ N_{\text{cum}}(D, t) = \int_{D}^{\infty} g(D') dD' \cdot f(t') dt' \]  
(B.1)

for an undisturbed geological unit a simple solution can be found:

\[ N_{\text{cum}}(D, t) = G(D) \cdot F(t), \]  
(B.2)

where \( G \) represents the cumulative crater size-frequency distribution and \( F \) the flux of projectiles.

In a unit where subsequent processes (erosion as well as deposition) occurred, the smaller crater population has been modified preferentially. Therefore, the measured crater size-frequency distribution will deviate from the crater-production function for craters smaller than a certain diameter \( d^* \). If the resurfacing event ended at a certain time \( t^* \), the surface accumulates craters as it is a fresh surface. In the cumulative description, the crater size-frequency distribution is represented by the sum of two (or more) populations, where the older one(s) do not cover the entire measurable size range. Such a distribution is represented by:

\[ N_{\text{eros}} = \int_{D_{\text{min}}}^{D_{\text{max}}} g(D') dD' \cdot \int_{0}^{t^*} f(t') dt' + \int_{D_{\text{min}}}^{D_{\text{max}}} g(D') dD' \cdot \int_{t^*}^{t_{\text{max}}} f(t') dt'. \]  
(B.3)

where the measurement is performed with a diameter range between \( D_{\text{min}} \leq d^* \leq D_{\text{max}} \) and the original surface has been formed at a time \( t_{\text{max}} \), the resurfacing ended at a time \( t^* \), with \( 0 \leq t^* < t_{\text{max}} \). If \( t^* = t_{\text{max}} \), no resurfacing occurred and the description follows Eqs. (B.1) and (B.2). On a planetary surface, a proper remote-sensing method to distinguish both populations is not known. Any attempts to distinguish between erosional state or other parameters are strongly misleading (and subjective). To understand the crater population accumulated on the surface after the resurfacing event, a method is needed to separate both populations. Due to the crater accumulation per time and the cumulative characteristics of the measured crater size-frequency distribution, the contribution of the old crater population has to be calculated with respect to their quantity in the diameter range larger than \( d^* \). The measured crater frequencies in the smaller crater size range, attributed to the resurfacing event, have to be reduced (by the number of craters produced in the period prior to the resurfacing event). Based on the known crater-production function, the cut-off diameter and the age of the old surface can be determined, while the exact resurfacing age remains obscured:

\[ N_{\text{eros}} = \int_{D_{\text{min}}}^{D_{\text{max}}} g(D') dD' \cdot \int_{0}^{t_{\text{max}}} f(t') dt' - \int_{D_{\text{min}}}^{D_{\text{max}}} g(D') dD' \cdot \int_{0}^{t^*} f(t') dt' \]

\[ + \int_{D_{\text{min}}}^{D_{\text{max}}} g(D') dD' \cdot \int_{t^*}^{t_{\text{max}}} f(t') dt'. \]

\[ = \left( G(D_{\text{max}}) - G(D^*) \right) \cdot F(t_{\text{max}}) \]

\[ - \left( G(D_{\text{max}}) - G(D^*) \right) \cdot F(t^*) \]

\[ + \left( G(D_{\text{max}}) - G(D_{\text{min}}) \right) \cdot F(t^*). \]  
(B.4)

Following the description in Eqs. (B.3) and (B.4), a fit of the crater production function \( G(D) \) allows one to predict the shape of the expected distribution in the large-size range of the “resurfacing population,” which gives a preliminary but overestimated age \( t_{\text{prelim}}^* \). Next, the expected number of craters accumulated on the younger surface is predicted. Thus, the contribution (in number) of the old crater population in the large branch can be calculated and the measured number of craters at the cut-off diameter \( d^* \) reduced, accordingly. By fitting the reduced distribution, the younger age is correctly determined. Equation (10) demonstrates that an iterative approach is necessary to determine the age(s) of resurfacing event(s) by fitting the well known crater production function to the crater size range smaller than a certain diameter \( d^* \) and to predict the expected number of craters for diameters larger than diameter \( d^* \). Reducing the measured number to the predicted one allows for the determination of the time when the resurfacing event ended, and therefore, the age of the younger surface. Resurfacing ages described in this work follow the approach described above, and for more details see Werner (2005).

Appendix C. Uncertainties in the relative and absolute ages

Prerequisite for the crater count statistics are (1) careful geologic mapping outlining homogeneous units, and (2) excluding sublimations, pits, volcanic and secondary craters. Contamination due to unrecognized global secondary craters unwittingly included in the measurements is less than 10% (old surfaces), and in most cases less than 5% (Werner, 2005; Werner et al., 2009). In recent years there has been an ongoing discussion (Hartmann, 2005) about the use of small craters in cratering statistics; specifically whether it is appropriate to count craters \( \leq 300 \) m in diameter because they might be distant secondary craters rather than primary craters (McEwen et al., 2005). In this work, none of the size-frequency curve measured show an deviation in steepness from the predicted crater size-frequency distributions, except for resurfacing `kinks', even down to crater sizes of about 100 m diameter (compare also Hartmann et al., 2008).

For relative ages (the crater frequency per unit area), the ages are different if the statistical error is not overlapping. For absolute ages, which are simply a translation between one number to another number based on the chronology model the same is valid. The general uncertainty of the applied chronology model is given by the uncertainties in the transfer between the lunar chronology model to Mars, and implies that either all ages are either too young or too old. However, Malin et al. (2006) estimated the current martian cratering rate, based on changes observed in MOC images over a seven-year period, and they found that the models that scale lunar cratering rates to Mars are consistent with the observed martian cratering rate (Hartmann, 2007; Hartmann et al., 2008).

References


