Mineralogy of recent volcanic plains in the Tharsis region, Mars, and implications for platy-ridged flow composition

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Volcanism on Mars is diverse and well developed, with evidence of abundant plain-style volcanism and a variety of edifices. At the scale of Viking imagery, early investigators observed that the most recent volcanism was located in distinct areas such as the Tharsis (including Olympus Mons) and Elysium regions (e.g., Greeley and Spudis, 1981). At that time, it was unclear if volcanic activity extended into the Late Amazonian (<500 My). A better picture of recent volcanism on Mars occurred until recently, but the mineralogy of recent lava plains is poorly known because few regions display fresh outcrops devoid of dust. Using visible and near infrared data of the Mars Express probe, two new volcanic plains in Noctis Labyrinthus have been identified, and the existence of a volcanic plain on the floor of Echus Chasma has been confirmed. Crater retention ages estimated for these three plains range between 50 and 100 My, corresponding to the Late Amazonian. These plains represent an excellent opportunity to constrain the mineralogy of recent volcanic rocks. Results show that basaltic compositions with plagioclase and high calcium pyroxene are predominant. The low olivine proportion suggests that the apparent fluidity of these flat plains is not related to magmas being ultramafic. In addition, a platy-ridged texture is observed in two of the studied regions. Our study shows, for the first time, that this texture is associated with volcanic rocks, and that these rocks are of typical basaltic mineralogy. Finally, these volcanic plains are located more than 1000 km east of previously known Late Amazonian volcanic centers of the Tharsis region, an observation to be taken into account when considering models of recent volcanism on Mars.

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1. Introduction
Volcanism on Mars is diverse and well developed, with evidence of abundant plain-style volcanism and a variety of edifices. At the scale of Viking images, early investigators observed that the most recent volcanism was located in distinct areas such as the Tharsis (including Olympus Mons) and Elysium regions (e.g., Greeley and Spudis, 1981). At that time, it was unclear if volcanic activity extended into the Late Amazonian (<500 My). A better picture of recent volcanism on Mars occurred until recently, but the mineralogy of recent lava plains is poorly known because few regions display fresh outcrops devoid of dust. Using visible and near infrared data of the Mars Express probe, two new volcanic plains in Noctis Labyrinthus have been identified, and the existence of a volcanic plain on the floor of Echus Chasma has been confirmed. Crater retention ages estimated for these three plains range between 50 and 100 My, corresponding to the Late Amazonian. These plains represent an excellent opportunity to constrain the mineralogy of recent volcanic rocks. Results show that basaltic compositions with plagioclase and high calcium pyroxene are predominant. The low olivine proportion suggests that the apparent fluidity of these flat plains is not related to magmas being ultramafic. In addition, a platy-ridged texture is observed in two of the studied regions. Our study shows, for the first time, that this texture is associated with volcanic rocks, and that these rocks are of typical basaltic mineralogy. Finally, these volcanic plains are located more than 1000 km east of previously known Late Amazonian volcanic centers of the Tharsis region, an observation to be taken into account when considering models of recent volcanism on Mars.

High resolution imagery and topography can be combined with visible and near infrared spectral data to study both morphology and composition of volcanic rocks on Mars (e.g., on Syrtis Major, Poulet et al., 2003). However, few spectral data exist for areas of recent volcanism, because such surfaces are often featureless due to widespread dust cover (e.g., Hamilton et al., 2003, Stockstill-Cahill et al., 2008). As described in this manuscript, three small regions (~50 km wide) were found to have formed recently (~100 My). Two of these areas were found in Noctis Labyrinthus, and one was found on the floor of Echus Chasma (Fig. 1). The resolution of OMEGA (Observatoire pour la Minéralogie, l’Eau, les Glaces et l’Activité) spectral data provides the possibility to analyze the composition of these recent plains in detail (Bibring et al., 2004).

Plain-style volcanism, corresponding to broad lava flows without obvious volcanic vents, is widespread on Mars (e.g., Hodges and Moore, 1994). However, few mineralogical data exist with which to constrain their composition independently from rheological evidence suggesting a high fluidity. Additionally, platy-ridged flows that are common in Martian plains are the subject of debate due to their morphologic similarity with potential ice rafts, suggesting frozen seas in the southern Elysium plains (Murray et al., 2005). Two of the three
 plains studied showed platy-ridged landforms with fresh outcrops enabling us to extract their mineralogy. Thus, our study focuses on the morphology, mineralogy of recent volcanic plains that represent an excellent opportunity to constrain the composition of recent volcanic rocks and understand their apparent fluidity, in order to provide insights for the origin of recent volcanism on Mars.

2. Methods

2.1. HRSC/Mars Express data

The HRSC camera acquired images in five panchromatic channels under different observation angles, as well as four color channels at a relatively high spatial resolution (Neukum and Jaumann, 2004). In our work we used only panchromatic nadir images, with a maximum spatial resolution from 10 to 40 m pixel$^{-1}$, and two panchromatic stereoscopic images with a spatial resolution usually degraded compared to the nadir image. The coordinates of ortho-rectified nadir images are defined in the planetocentric system of the Mars IAU 2000 ellipsoid (Seidelmann et al., 2002). The following images have been processed: #1977, #1999, #2402, #2479, and #3155 in the Noctis region, with a spatial sampling of the nadir image at 15 m pixel$^{-1}$ for #1977, #1999, and #3155, and 40 m pixel$^{-1}$ for the #2402 and #2479. Images of orbit #71 and #5379 are used in the Echus Chasma region with a spatial sampling at 30 m pixel$^{-1}$.

HRSC DEMs (Digital Elevation Models) were computed using the photogrammetric software developed at the DLR and the Technical University of Berlin (Scholten et al., 2005; Gwinner et al., 2005, Ansan et al., 2008). The image correlation was performed using a matching process at a different spatial grid size (Scholten et al., 2005). The height was calculated taking into account the Martian geoid, defined as the topographic reference for Martian heights, i.e. the areoid (Smith et al., 1999). In the Noctis Labyrinthus region, the DEMs were constructed using spatial gridding of ~30 m/pixel for orbits #1977 and #1999, and 100 m for orbit #2479. In the Echus Chasma region, only the MOLA (Mars Observer Laser Altimeter) DEM at 1/128 degree was used for the topography.

2.2. OMEGA/Mars Express data

OMEGA is a visible and near infrared (VNIR) hyperspectral imager providing three-dimensional data cubes with spatial samplings from a few kilometers to 300 m. For each pixel it provides spectra between 0.35 and 5.1 µm, with 352 contiguous spectral elements (spectels), 7–20 nm wide. The spectrometer consists of three detectors (from 0.35 to 1 µm, from 0.9 to 2.7 µm, and from 2.5 to 5.1 µm) (Bibring et al., 2004). For this study we used data recorded by the second detector where the signatures of the mafic minerals are well characterized. Pyroxene is detected from the presence of a broad band at 1.9 and 2.3 µm, respectively, for low-calcium pyroxene (LCP) and high calcium pyroxene (HCP). In our study, the detection of pyroxene is done with the spectral index used by Poulet et al. (2007) sensitive to the presence of both HCP and LCP. To account for instrumental and atmospheric biases, the detection is considered positive for values of the pyroxene spectral parameter larger than 1% (Poulet et al., 2007). The detection of olivine is based on the broad 1 µm band from the spectral parameter defined in Mustard et al. (2005), but no olivine has been detected by this method on the region studied. The depths of absorption bands were determined and mapped, allowing the presence of minerals to be identified in the top few micrometers of the Martian surface.

The identification of minerals from band depths does not, however, provide a direct determination of all minerals. In addition, the spectral parameter used for pyroxene is not designed to discriminate between the LCP end-member and the HCP end-member. Modeling of NIR spectra has been demonstrated to provide an accurate estimate for mineral abundances in mixtures of basaltic granular materials for a wide range of particle sizes (Poulet and Erard, 2004). As described by Poulet et al. (2009a), scattering models may be used to explore the parameter space and the influence of grain size of mineralogical members, including spectrally neutral components. The model has two free parameters for each end-member: the average grain size and the relative abundance. Since aerosols influence spectral slopes, one additional free parameter is used to adjust the spectral slope continuum. The spectra are fitted in the 0.99–2.49 µm wavelength range (OMEGA SWIR-C channel) using a simplex minimization algorithm. The synthetic spectrum is a nonlinear combination of the optical indices of the minerals selected to be part of the mixture, in proportion to their abundance. Upon obtaining a data spectrum fit, the algorithm supplies the user with a model-derived spectrum as well as with a percentage and grain size for each end-member used in the fit. The principal mineralogical characteristics of the three areas studied were determined using this method and are summarized in Table 1.

<table>
<thead>
<tr>
<th>Location</th>
<th>HCP</th>
<th>LCP</th>
<th>LCP/HCP</th>
<th>Neutral</th>
<th>Olivine</th>
<th>Dust</th>
<th>RMS</th>
</tr>
</thead>
<tbody>
<tr>
<td>Echus Chasma</td>
<td>25 ± 6</td>
<td>6 ± 3</td>
<td>23 ± 13</td>
<td>47 ± 10</td>
<td>&lt;5</td>
<td>17 ± 13</td>
<td>0.26</td>
</tr>
<tr>
<td>Noctis Labyrinthus 1</td>
<td>34 ± 5</td>
<td>5 ± 3</td>
<td>17 ± 10</td>
<td>60 ± 10</td>
<td>&lt;5</td>
<td>&lt;1</td>
<td>0.39</td>
</tr>
<tr>
<td>Noctis Labyrinthus 2</td>
<td>30 ± 5</td>
<td>8 ± 3</td>
<td>29 ± 15</td>
<td>48 ± 7</td>
<td>&lt;5</td>
<td>12 ± 9</td>
<td>0.39</td>
</tr>
</tbody>
</table>

Values are percentage (including the LCP/HCP ratio) and standard errors for abundances indicate ±1σ. Neutral end-member corresponds to plagioclases.
2.3. Other datasets

Mars Observer Camera (MOC) images are used to complement HRSC nadir images at a better resolution, typically 2.8 m pixel\(^{-1}\) (Malin et al., 1992). Nighttime and daytime infrared images of the Thermal Emission Imaging System (THEMIS) onboard Mars Odyssey were used to provide a view of the surface properties (Christensen et al., 2003). Only the mosaics of thermal images are used at 230 m pixel\(^{-1}\). Wechos et al. superimposed these images and superposed them onto HRSC or THEMIS daytime images in order to observe where thermal images may be indicative of the presence of indurated rock outcrops or coarse grains (as seen in red for high nighttime temperatures) compared to more dusty, less indurated particles (as seen in blue for low night time temperatures). Values of thermal inertia for the regions of interest have been extracted from the TES (Thermal Emission Spectrometer) thermal inertia map at a resolution of 3 km (Putzig et al., 2005).

3. The floor of Noctis Labyrinthus Chasmata

3.1. Regional context and the distribution of pyroxenes as seen by OMEGA

Pyroxene signatures are present throughout all of the southern part of Tharsis and in Valles Marineris (Fig. 2). Hence, these regions are devoid of dust as seen on the thermal inertia map (Fig. 1). However, most pyroxene signatures are observed either on dark sand dunes filling canyons floor, such as in Candor Chasma (Mangold et al., 2008), or on canyon walls (e.g. Mustard et al., 2005). The source of material is unknown for canyon floors. For canyon walls, rocks correspond to thick accumulations of lava flows of Hesperian or Noachian age, which are beyond the scope of our study. By examining the region of Noctis Labyrinthus (Fig. 2), we observe pronounced pyroxene signatures in the west of Syria Planum and West Tithonium Chasma.

Syria Planum is a volcanic plain with many small low shield volcanoes of Hesperian age (Baptista et al., 2008). To the east, West Tithonium Chasma is covered extensively by sand sheets and dunes that bury the canyon, and part of the Oudemans crater (the 100 km diameter crater to the southeast of the canyon). Therefore, none of these signatures have been found to be correlated with young volcanic rocks. However, our interest focused on two spots (NL1 and NL2 on Fig. 2) inside Noctis Labyrinthus (Fig. 2), where pyroxene signatures appear to be pronounced and the thermal inertia is high. Spectra of both regions (Fig. 3) fit well with typical laboratory spectra of pyroxenes, especially HCP, with broad 1 µm and 2.3 µm bands.

3.2. West canyon floor (NL1)

The first canyon floor analyzed, NL1, is located at 99W, 7S (Fig. 4). It is about 50 km from west to east, 60 km from north to south, and 5 km deep below plateau level. The HRSC image shows a smooth floor only interrupted by small buttes. The albedo of the floor is very low compared to walls and plateaus (Fig. 4a), and low albedo correlates with the high night time temperatures seen on the floor (Fig. 4d), corresponding to areas with more than 500 J m\(^{-2}\) K\(^{-1}\) s\(^{-1/2}\) on the TES thermal inertia map (Putzig et al., 2005) (Fig. 1b). Only the uppermost part of the canyon walls (rock scarps) display nighttime temperatures as high as the canyon floor (Fig. 4d). Pyroxene signatures detected by OMEGA are found only on the canyon floor. The topography of the floor measured by MOLA data (Fig. 4b,c) indicates that the floor is at a constant altitude of ~2250 m with limited variation (< 50 m). All of the pyroxene signatures (Fig. 4e) are detected beneath this elevation contour.

A closer look to the east of this canyon is possible with more resolved HRSC data from orbit 1999 (Fig. 5). Here, the HRSC DEM allows us to observe that all small buttes located on the canyon floors are located above the 2250 m contour. This contour separates that part of the floor with low albedo (0.15 at 1 µm OMEGA reflectance)
relative to surrounding bright terrain of smooth texture, containing pyroxene. These observations confirm that the whole floor is very flat at the elevation of 2200 to 2250 m over the entire canyon floor area.

Close-ups of the canyon floor reveal further landforms not visible at a broader scale. First, a circular shape 1 km in diameter and about 100 m in depth is observed beyond the southern edge of the plain (Fig. 6a,b). It is breached in its northern part by a 200 m large, 30 m deep channel which joins the canyon floor 5 km to the north and vanishes inside the canyon floor shortly after intersecting the 2250 m elevation contour (Fig. 6b). In addition, most of the canyon floor is smooth at the HRSC scale with only a few sand dunes present in the central part. Nevertheless, the MOC close-up shows small impact craters and local outcrops, especially around unfilled pits (Fig. 6c), suggesting that sand exists but does not mantle all the surface. These images also display a lineation of NW–SE direction, resembling a buried dyke or a fracture zone, with collapse pits 100 m wide sitting on the fracture (Fig. 6c). Finally, the area located east of the canyon floor exhibits not only small buttes, but a series of pitted cones of about 100 m in diameter, which are observed at the contact of the 2250 m altitude contour (Fig. 6d).

All these observations listed above indicate a volcanic origin for the canyon floor. By itself, the pyroxene-rich mineralogy coupled with the low albedo and the strong thermal inertia is good evidence for a volcanic origin (Fig. 4). Additionally, the combination of the deep circular landform together with a channel (Fig. 6) is similar to terrestrial volcanic vents with lava channels emerging from them (e.g. pit crater with lava channel in the Snake River plain, Idaho, USA, Fig. 5G, in Hodges and Moore, 1994). Pitted fractures inside the canyon floor might correspond to other vents, which were partially filled by the lavas they extracted. The pitted cones to the east might be an additional source of lavas. Alternatively, they might correspond to rootless cones, also named pseudocraters. Rootless cones form at the contact of lava flows and a volatile such as liquid water or water ice (e.g. Lanagan et al., 2001). Their presence at the contact with the
canyon floor is consistent with such a hypothesis. The apparent size of these cones do match terrestrial cones as well as some other Martian examples (Lanagan et al., 2001; Meresse et al., 2008) and may indicate the presence of liquid water or water ice at the time of the volcanic activity. Finally, the constant altitude of the canyon floor suggests that the volcanic material is fluid enough to form a smooth plain which follows an equipotential surface. All this evidence favors formation as a lava plain, covering about 450 km² if the 2250 m contour is taken as a reference.

3.3. East canyon floor

The second canyon floor analyzed, NL2, is located at 96.2W, 7.2S (Fig. 7). It is about 50 km in length from west to east, 40 km in length from north to south, and 6 km in depth. The HRSC image shows a smooth floor with small buttes in the eastern and southern part. The albedo of the floor is low when compared with the walls and plateaus (0.11 at 1 µm OMEGA refection), and displays a gradation through the floor from the west to the east, the eastern region being darker than the western region (Fig. 7a). Low albedo correlates with the high night time temperatures seen on the canyon floor (Fig. 7d), reaching up to 600 J m⁻² K⁻¹ s⁻¹/₂ on TES data in the darkest areas (Putzig et al., 2005) (Fig. 1b). Here too, only the uppermost part of canyon walls (namely the rock scarps) display temperatures as high as the canyon floor. Pyroxene signatures detected by OMEGA are found mainly on the eastern part of the canyon floor (Fig. 7e).

The topography of the floor as measured by HRSC data (Fig. 7b,c) indicates that the floor is constrained between the altitudes of 1850 and 2000 m. The 2000 m contour follows the smooth texture of the canyon floor very well. We observe that the transition between high temperatures (red) and low temperatures (blue to yellow) at the southern and northern edge of the canyon floor correlates with the 2000 m altitude level (Fig. 7d). Dust, as seen from blue colors (low night temperatures) and high albedos in the HRSC image, mantles most of the canyon walls and debris aprons. As dust deposition is a gradual process at the regional scale, this observation likely illustrates that dust was deposited before the canyon floor was resurfaced, and that dust was buried beneath the high thermal inertia material. Even so, the fact that the albedo is slightly higher in the western part of the canyon floor suggests that the western part has been slightly mantled by a thin (<10 cm) coating of fresh dust hiding the rock surface at OMEGA wavelengths. An elliptical feature, marked by “E” is visible in the albedo (Fig. 7a), in the topography (Fig. 7c), and in the temperature maps (Fig. 7d). This landform is oriented NW–SE, is a few tens of meters deeper than the rest of the canyon floor, and could have served as a trap for recent dust deposition.

A close-up on HRSC and MOC data shows a set of different textures (Fig. 8). First, the southern part exhibits a specific texture that differs from the smooth canyon floor. Here, plains appear as if it has been disrupted by liquefaction, or some kind of fluid activity. Indeed, small pieces of plains are found in the middle of this area and resemble solid plates floating over a fluid medium (Fig. 8a). The plates are still as smooth as the plains material, as seen on the MOC close-up (Fig. 8b). In between these plates, the texture is rough with a lot of fractures that do not cross the plates. Plates are up to 1 km long and are less than 30 m thick because the resolution of the HRSC DEM is insufficient to detect the plates.

Outside this area, plains exhibit frequent small ridges (marked by “R” in Fig. 8c) several kilometers long and approximately 100 m wide. The ridges display a positive relief, as if they were formed by compression in soft material, but their curved shape without straight faults (Fig. 8c) does not plead for a tectonic origin by brittle faulting. In Fig. 8c, left of the ridge, curvilinear shapes of material that appear to float over a fluid medium (Fig. 8a). The plates are still as smooth as the plains material, as seen on the MOC close-up (Fig. 8b). In between these plates, the texture is rough with a lot of fractures that do not cross the plates. Plates are up to 1 km long and are less than 30 m thick because the resolution of the HRSC DEM is insufficient to detect the plates.
have been deformed in a ductile way can be observed. Small impact craters can also be observed in this image.

In summary, this volcanic plain has a relatively high nighttime temperature with rough textures of bedrock exposure with low albedos and pyroxene signatures on fresh outcrops. These characteristics argue in favor of a volcanic origin such as lava flood plains. The plate shape, size and width are consistent with those frequently present in Martian lava flood plains, usually referred to as platy-ridged lava flows (e.g., Kesztölyi et al., 2000). Other landforms such as ridges and curved textures are also consistent with viscous lava flows, despite the fact that this type of ridge does not seem frequent in other Martian flood plains. The total surface of this volcanic plain is about 800 km². A difference between the NL2 and NL1 plains is that there is no obvious vent found in, or along the plain NL2. Nevertheless, it is possible that the elliptical trough (E) corresponds to a lava vent that has been entirely buried beneath its flows.

4. Echus Chasma floor

Echus Chasma is the source area of the Kasei Vallis outflow channels that dissect volcanic plateaus from the equator to 20N, west of the large volcanoes of Tharsis (Fig. 1). The Echus-Kasei valley is partially filled by lava flows originating from the west, from the main volcanic center (e.g. Scott and Tanaka, 1986). While no vent is observed in our studied area, volcanic fissures exist on the floor of the valley north of the area studied, suggesting that local emissions from the Chasma floor are also possible (Chapman et al., 2010-this issue).

The studied area is a zone with two types of material, a high thermal inertia, low albbedo region, in the middle of plains that are generally of low thermal inertia and high albbedo (Fig. 9). Pyroxene is detected where the albbedo is low (0.19 in 1 µm OMEGA reflectance) and the inertia high (Fig. 9d, e). The morphology observed in these plains shows a lack of sand dunes and a rough surface at MOC scale suggesting the surface is relatively free of aeolian material (Fig. 9a,b,f). Typical spectra of this region have a well defined pyroxene signature, as visible in Fig. 3. In contrast, the high albbedo region in the surroundings consists of low inertia material that appears featureless in OMEGA spectra. It most likely corresponds to rocks mantled by eolian deposits such as bright dust. The window of low albedo rocks therefore signals a location where the spectral composition can be explored. The high albedo (0.19) compared to the Noctis plains (below 0.15) could suggest that a thin dust mantle (<100 µm thick) is present, but not in sufficient proportion to mask the detection of pyroxene.

The overall topography of the Echus Chasma floor is flat, suggesting that a low viscosity fluid filled the trough (Fig. 9c). At the scale of MOC images, surface morphology is complex and includes pressure ridges, straight fractures, and rafted plates (Fig. 9a,b). Individual plates are several tens of meters to several kilometers wide, relatively smooth, and separated by rougher and brighter terrains. The smooth plates often seem to fit together like a jigsaw puzzle. Therefore, these plates can be interpreted to be pieces of a ruptured crust that have rafted over a fluid medium. Pressure ridges are interpreted to exist at the edges of plates, since these plates are rapped into each other during emplacement. These characteristics are typical of platy-ridged flows, as described by Kesztölyi et al. (2000). In summary, this plain displays mineralogical features and morphologic patterns that correspond to a volcanic filling. The rough texture observed on MOC images, the lack of sand dunes, and the relatively high thermal inertia favor a rocky bedrock locally exposed without a thick dust mantle. The area of the plains defined by the OMEGA signature is approximately 2000 km².

5. Determination of ages and compositions

5.1. Ages

Ages are obtained counting craters in the three regions studied according to the diagram and methods described in Hartmann and Neukum (2001). This diagram divides the distribution range (crater diameter) into log intervals incremented in square root 2. Using studies from the Moon, meteoroid distributions and impact frequency, predicted “isochrons,” or crater size-frequency distributions, have been derived for well-preserved surfaces of various ages, such as 1 Ga, 100 Ma, 10 Ma, and so on. Determined ages are valid for fresh surfaces.
not affected by erosion or deposition. Crater counts should follow isochrons when the surface is fresh, while crater counts crossing isochrons indicate a surface that has been subsequently modified. Counts were done in a GIS software using sinusoidal projections of HRSC and MOC images for measuring crater diameters and the area of the unit.

NL1 crater counts are determined for over 250 km² on the eastern part of the plain (Fig. 5). We limit counts to HRSC data due to the paucity of images with better resolution in this case. An age following approximately the 100 My isochron can be retrieved graphically. A model age of 77 ± 21 My is determined using the more precise mean square root method proposed by Vaucher et al. (2009). The younger age with this method is due to the fact that smaller craters are more numerous, therefore statistically more reliable in the mean square method.

NL2 crater counts are determined over an area of 800 km² from HRSC images, and over an area of 210 km² from the three MOC images crossing this part of the plain. The plot reveals a good consistency between craters counted in HRSC and MOC images (Fig. 10, circles). The isochron for this plot corresponds to an age of 50 to 100 My. This age is well constrained from the large range of craters used. A precise model age of 66 ± 5 My is determined using the method of Vaucher et al., 2009.

Echus Chasma Floor crater counts were derived from the relevant HRSC mosaic over an area of 1600 km². The crater counts are close to the 100 My isochron (Fig. 10, triangles). The slight shift to a younger age for smaller craters could be related to the statistically low number of large craters, or to erosion/deposition processes that influenced part of the canyon floor, thus removing some small craters. Despite this slight difference in the isochron slope, ages in the range 50–100 My provide reasonable fits. A model age of 52 ± 8 My is determined using the mean root squares method (Vaucher et al., 2009), assuming the smaller bin is not affected by more recent obliteration. This age is consistent with that of about 70 My established using cumulative methods (Chapman et al., 2007, 2010-this issue).

Plots of crater density follow isochrons well enough to confirm that the ages correspond to the formation of the plains, or the plots would have crossed isochrons without following them. One should remember that the absolute ages have uncertainties of a factor 3 to 4 (e.g. Hartmann and Neukum, 2001). In addition, the correction factor due to a decreasing impact flux through time might underestimate these ages by a factor of roughly 3 (Quantin et al., 2007). Even with these limitations, the three studied plains are still extremely young relative to the majority of the Martian surface (300 My is an age well into the Late Amazonian epoch). The estimated ages can be compared to other studies of volcanic centers on Mars. These ages are among the youngest found on Mars for volcanic plains, and are of the same order as those proposed for the calderas of Arsia Mons or Olympus Mons in the Tharsis region (e.g., Neukum et al., 2004) and volcanic plains in Central Elysium Planitia (Vaucher et al., 2009).

5.2. Composition

The mineralogy derived from spectra is shown in Table 1. Plagioclases abundance, inferred by OMEGA is included in a mineral group referred to as “neutral components,” for which the abundance is the combined abundance of spectrally featureless phases in the near infrared (plagioclases, high silica phases and quartz). Plagioclases are expected to be predominant in this group for smooth lava plains.

Grain sizes were determined in models from a few tens to a few hundreds of micrometers, except for dust which was below 10 μm. Results indicate that plagioclase, HCP and LCP are the dominant minerals for the three regions listed in Table 1. A contribution of dust is required in two regions. Olivine was not required in the fit of spectra. This shows that olivine was below the detection limit (~5% for grain size of 100 μm, or a slightly higher proportion for grains below 10 μm). The derived abundances of plagioclases and pyroxenes (40 to 60% plagioclases, 30 to 40% pyroxenes) are typical of basaltic
composition. The three regions also display a very low LCP/HCP ratio, some of the lowest ratios known to date on Mars. Indeed, most Noachian crustal rocks or Hesperian aged plains have LCP/HCP ratios of 1:1 to 1:4 (Poulet et al., 2009b), whereas NL1 reaches a ratio of 1:6 (17% in Table 1). Such characteristics were interpreted to support the idea that there is a change in the LCP/HCP ratio through Martian
recent volcanism in the Tharsis region is generally considered to be limited to the main volcanic edifices, such as Arsia or Olympus Mons and surrounding plains. In our study, we have found two previously undescribed plains <100 My old. These plains were previously mapped as undifferentiated floor material of Early to Middle Amazonian age and of diverse nature (Scott and Tanaka, 1986). No recent volcanism has ever been reported inside the Valles Marineris–Noctis Labyrinthus troughs, despite the large amount of Hesperian age volcanism surrounding these areas (e.g., Baptista et al., 2008). In addition, Echus Chasma floor has been confirmed to be covered by volcanic terrains of similar ages. The location of the three studied regions, much further to the East than the main volcanic centers of Tharsis, is a feature of interest in itself to understand the evolution of Valles Marineris Chasmata.

The Echus Chasma floor could have been filled from lava flows coming from the western plateau of the Echus–Kasei system of troughs, as suggested by Viking based mapping (Scott and Tanaka, 1986). Such an observation would suggest that the volcanic centers are actually on the Tharsis plateau, close to the volcanoes. Nevertheless, fractures and landforms observed on the bottom of Echus Chasma north of the studied area suggest that lavas were also extruded from fissures on the canyon bottom (see Chapman et al., 2007, 2010-this issue for more details), not only from the western plateaus. This suggests that local magmatic sources also exist below the canyon floor. The two Chasmata of Noctis Labyrinthus were not filled from the plateaus since there is no sign of lavas flowing into the canyons. NL1 displays volcanic vents, thus showing that the origin of the volcanism is local, below the canyon floor.

As the origin of the volcanism is not the focus of our study, we do not enter into in-depth discussions on that question, but the observed volcanic activity, well to the east of the Tharsis area, should be taken into account in geophysical models attempting to explain recent volcanism on Mars (e.g., Schumacher and Breuer, 2007). Possible explanations include the strong faulting in this region, especially in Noctis Labyrinthus, compared to regions devoid of widespread faulting, and, the thinner crust on the floor of Chasmata than on the plateau (>5 km of difference of elevation). Both of these features may enable magmatic activity to reach the surface.

6.2. The composition of platy-ridged flows

The existence of platy-ridged flows with rafted slabs and pressure ridges is well known on Mars (i.e., Keszthelyi et al., 2000). However, this type of pattern is not frequent for terrestrial volcanism, and has been discussed almost exclusively for flows in Iceland (Keszthelyi et al., 2004, Haack et al., 2006). Therefore, not all workers agree that these features on Mars are lava flows. Mudflows and pack ice (icebergs floating over water) have a very similar shape, with rafted slabs of ice over water or rocks over mud. Such a scenario has been proposed for platy flows seen in Kasei Vallis by Woodworth-Lynas and Guigné (2003) and Williams and Malin (2004). In this respect we note that Kasei Vallis originates in Echus Chasma where our platy flows are observed. In addition, Central Elysium Planitia plains have been proposed to have been filled by a frozen sea displaying rafted plates (Murray et al., 2005). The problem of these contradictory interpretations is that they are based on geomorphic landforms of very similar shape. No spectral study of platy-ridged flows has been performed to discriminate these hypotheses, principally because all of these regions are blanketed by dust. However, in the regions studied here, we observe two plains that display platy-ridged textures: the Eastern Noctis Chasma (NL2) and Echus Chasma (Figs. 8 and 9 respectively). Both of these areas display a mineralogy of basaltic lava flows with predominant plagioclases and pyroxenes. While we cannot confirm that all platy-ridge flows are magmatic in origin, our data provide the first unambiguous evidence that such morphological features on Mars may indeed be associated with basaltic lava flows.

Models of platy-ridged lava flows generally show a viscosity in the range of 100–1000 Pa s. (Keszthelyi et al., 2000). Viscosities of below 10^3 Pa s, and yield strengths less than 200 Pa are found in the Central Elysium Planitia (Vaucher et al., 2009), a location where extensive platy flows have also been reported (Keszthelyi et al., 2000). Unusually low viscosity or low yield strengths for silicate magma may be related to different parameters, including a high temperature or a low proportion of Si. The mean compositions of the two plains are that of classical basalts, and not that of an ultrabasic rock. Although the mineralogy of platy-ridged flows has been established at the surface of the flows, this does not give the mineralogy at the bottom of flows. A downward segregation of olivine may have taken place, thus being a possible explanation for the lack of olivine detection in the OMEGA spectra. However, the fact that olivine is systematically absent, or at least in limited proportion in small grains, even in eroded parts of NL2, and that the detection of pyroxene, another dense mineral, is clear, are both arguments that plead against a loss of olivine from the surface. Thus, the lack of detection of olivine suggests that an ultrabasic composition was not necessary to explain the lavas fluidity. A similar conclusion has been drawn from Icelandic examples where platy flows correspond to tholeiitic basalts (Keszthelyi et al., 2004). Thus, mechanisms such as deep crustal melting, dissolved volatiles (Vaucher et al., 2009), or high effusion rates (Keszthelyi et al., 2000) are necessary to explain the fluidity of platy-ridged flows rather than an ultramafic composition.
6.3. The composition of young lava flows compared to SNC meteorites

Shergottite meteorites generally accepted to be of Martian origin are commonly dated at ages between 175 and 475 Ma (e.g., Nyquist et al., 2001), a range comparable to the youngest volcanic areas on Mars. However, the link between basaltic shergottites, those expected to outcrop near the surface, and Martian volcanism as observed from orbit has not been very successful, especially due to the extensive dust mantle overlying young volcanic regions (e.g., Hamilton et al., 2003). Since the mineralogy of rocks from older regions does not fit the mineralogy of SNC meteorites, a conclusion has emerged that shergottite compositions are not found in spectral data because the youngest lava flows cannot be analyzed (Hamilton et al., 2003). In our study, we identified and determined the mineralogy of three areas of non mantled lava flows from the Late Amazonian, with ages <100 My. Determined ages are slightly younger than the youngest ages usually considered for shergottites, but these ages are not significantly different due to the uncertainties associated with the determination of absolute ages using the crater count method. Our results indicate that the mineralogy of these three lava plains studied is also consistent with the mineralogy of basaltic shergottites in terms of LCP/HCP ratios. Indeed, most of the basaltic shergottites are richer in LCP when compared to the proportions derived from the relevant OMEGA data. For example, data for Shergottites suggest LCP/HCP ratios from 1:1 to 18:1 (McSween and Jarosewich, 1983; Barrat et al., 2002; Taylor et al., 2002, Hamilton et al., 2003). In contrast, our studied areas have average LCP/HCP ratios from 1:4 to 1:6. Other regions of recent volcanic activity may have different ratios closer to those observed in SNCs. Nevertheless, shergottites ages of approximately 4.0 Gy determined by Bouvier et al. (2005) may be an alternative explanation. If this is the case, the observed discrepancy is explained because shergottites would not be relevant for a comparison with recent volcanism.

7. Conclusion

We analyzed the morphology and mineralogy of some of the most recent volcanic areas in the East Tharsis region of Mars. Findings include the following:

(i) Previously unknown volcanic plains 50 to 100 My old are found in two canyon floors of Noctis Labyrinthus, while similar recent ages were confirmed for the floor of Echus Chasma.

(ii) These volcanic plains are located >1000 km east of previously known Late Amazonian volcanic centers of the Tharsis region, an observation to include in models of recent Mars volcanism.

(iii) The platy-ridged textures observed in two of the studied regions are associated with volcanic lava flows as seen from the basaltic mineralogy coupled with volcanic landforms.

(iv) The apparent low viscosity of platy lava flows may be due to parameters other than low Si content of ultramafic rocks because of the limited olivine content (~5% assuming ~100 µm grains) and the classical basaltic mineralogy obtained.

(v) The low-calcium/high calcium pyroxene ratios found by modeling for these young lava flows are different from that of putative young Martian meteorites (basaltic shergottites).

Finally, our study shows that in the absence of a widespread recent region devoid of dust, we need to focus on small dust-free windows to characterize the mineralogy of recent volcanic on Mars.

References


