A Late Amazonian alteration layer related to local volcanism on Mars


Abstract

Hydrated minerals on Mars are most commonly found in ancient terrains dating to the first billion years of the planet’s evolution. Here we discuss the identification of a hydrated light-toned rock unit present in one Chasma of the Noctis Labyrinthus region. Stratigraphy and topography show that this alteration layer is part of a unit that drapes pre-existing bedrock. CRISM spectral data show that the unit contains hydrated minerals indicative of aqueous alteration. Potential minerals include sulfates such as bassanite (CaSO$_4$.1/2H$_2$O) or possibly hydrated chloride salts. The proximity of a smooth volcanic plain and the similar crater model age (Late Amazonian, <100 Myr) of this plain and the draping deposits suggest that the alteration layer may be formed by the interaction of water with ash layers deposited during this geologically recent volcanic activity. The alteration phases may have formed due to the presence of snow in contact with hot ash, or eventually solid–gas interactions due to the volcanic activity. The relatively young age of the volcanic plain implies that recent alteration processes have occurred on Mars in relation with volcanic activity, but such local processes do not require conditions different than the current climate.

1. Introduction

The OMEGA (Observatoire pour la Minéralogie, l’Eau, les Glaces et l’Activité) spectrometer on the Mars Express orbiter has detected sulfates in the layered deposits of Valles Marineris, in the Margaritifer Terra chaos region, and the Meridiani Planum region (Gendrin et al., 2005; Arvidson et al., 2005). The strong correlation of all sulfate signatures with light-toned layered deposits over large regions shows that their formation process is not restricted to highly localized aequous activity. These units might signify a specific period of sulfate formation during the Late Noachian to Late Hesperian after a period dominated by formation of phyllosilicates (Bibring et al., 2006). For example, phyllosilicates found in the Late Hesperian after a period dominated by formation of phyllosilicates in the Nili Fossae region are part of massive material of the highland crust in which clearly stratified units exist only locally as buried material (Mangold et al., 2007), whereas sulfates elsewhere on Mars are uniquely present as stratified deposits (Gendrin et al., 2005; Arvidson et al., 2005; Bibring et al., 2007; Mangold et al., 2008a; Le Deit et al., 2008).

Sulfates have been interpreted to form on Mars by chemical precipitation through evaporitic processes or alteration through groundwater circulation (Gendrin et al., 2005), as volcanic ashes altered by fumaroles (McCollom and Hynek, 2005), or glacial sediments reworked by eolian processes (Niles and Michalski, 2009). Thus, many issues remain open concerning their formation. One key question is the role of volcanism in the production of sulfur and its oxidation into sulfates. Currently no clear relationships have been established between the broad sulfate-bearing deposits such as in Terra Meridiani and volcanic landforms of similar age. Putative ash deposits were proposed to explain part of the sulfate-bearing outcrops (McCollom and Hynek, 2005), but no obvious volcanic landform has been observed to confirm this hypothesis. The sedimentary cross-bedding and Br/Cl variations with stratigraphy observed at the Opportunity landing site are better explained by deposition in a shallow water environment and eolian transport (Squyres et al., 2006; McLennan et al., 2005).

In this study, we focus in one of the Chasma of the Noctis Labyrinthus region (Fig. 1). A Late Amazonian volcanic plain covers the bottom of this Chasma (Mangold et al., 2009) and can be used to...
determine relative ages of units once stratigraphic contacts are established. A series of hills at the southeastern margin of this plain exhibit a light-toned layer. We use high resolution images and topography to show that this light-toned layer is at the base of a sequence of fine layered materials that drapes pre-existing bedrock. Using Compact Reconnaissance Imaging Spectrometer for Mars (CRISM) spectra (Murchie et al., 2007), we show that this light-toned layer contains hydrated sulfates or chloride salts. Crater counts of the different units show that both the volcanic plain and the layered deposits are Late Amazonian.

2. Data

The High Resolution Stereo Camera (HRSC) instrument acquires images in five panchromatic channels under different observation angles, as well as four color channels at a relatively high spatial resolution (Neukum et al., 2004). In this work we used only panchromatic nadir images, with a maximum spatial scale 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 and #2402 with a spatial sampling of the nadir image of 15 m pixel\(^{-1}\) and 40 m pixel\(^{-1}\) respectively. Further, a HRSC Digital Elevation Models (DEM) of #1977 was computed using the photogrammetric software developed at the DLR and the Technical University of Berlin (Scholten et al., 2005; Gwinner et al., 2007; 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 (Smith et al., 1999). The final DEM was constructed with a spatial grid of 30 m High Resolution Imaging System (HiRISE) images were used to focus on the detailed morphology of the studied area (McEwen et al., 2007) and combined with the HRSC images and DEM into a geographic system.

Mineralogy was determined from data acquired by the CRISM spectrometer, which is a visible and near-infrared (VNIR) hyperspectral imager that measures reflected sunlight from 0.4 to 4.0 \(\mu\)m in 544 channels at a spatial scale of \(\sim\)18 m pixel\(^{-1}\) (Murchie et al., 2007). Our CRISM and OMEGA data analysis focused on the detection of hydrated minerals using the 1.9 \(\mu\)m \(\text{H}_2\text{O}\) combination bend-plus-stretch absorption band. Many hydrated mineral phases, including sulfates and phyllosilicates, exhibit an absorption at 1.9 \(\mu\)m due to \(\text{H}_2\text{O}\). OMEGA data also indicated the presence of pyroxenes on the Chasma floor (Mangold et al., 2009). The regions

Fig. 1. Geologic context of the studied area. (a) MOLA map of Noctis Labyrinthus located between Valles Marineris and the Tharsis volcanoes. (b) HRSC image and DEM (in color and 250 m spaced contours). The studied area is located on the floor of a 50 km large Chasma. (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)
with hydrated mineralogy observed in CRISM data were not detected with OMEGA due to their small size and the lower spatial resolution of the OMEGA data.

3. Observations

3.1. Geologic context and main units

Noctis Labyrinthus is located west of the Valles Marineris canyon system. OMEGA and HRSC data have been used to identify two recent volcanic plains units in depressions in Noctis Labyrinthus (Mangold et al., 2009). Several Mars Observer Camera (MOC) images of these depressions display enigmatic light-toned layering with unknown composition (examples: MOC#R0900388, E0101736). Elsewhere in Valles Marineris, light-toned layers are often associated with the presence of hydrated sulfates (Gendrin et al., 2005). Unlike in the neighboring chasmata (Tithonium, Ius, and Melas), OMEGA detected no widespread sulfate-bearing deposits in Noctis Labyrinthus. However, higher-resolution CRISM data show that different troughs in Noctis Labyrinthus have interesting materials such as phyllosilicates, sulfates and hydrated silica (Milliken, 2008; Mangold et al., 2008b).

Fig. 2 shows the smooth and flat volcanic plain (VP) located in the west portion of the Chasma, located at (96°W, 7°S), which has a low albedo and a basaltic composition (Mangold et al., 2009). A few sand dunes are present here, but most of the plain corresponds to a high thermal inertia surface with a rough texture showing that this plain is a lava plain (see details in Mangold et al. (2009)). The lava plain embays and onlaps nearby buttes and is thus younger and stratigraphically higher than the buttes, as visible in the upper left part of Fig. 2. To the east, many buttes are present at the foot of the canyon walls (Figs. 1 and 2). A smooth and flat surface, brighter than the lava plain, is identified by the letter M in Fig. 2. This unit appears more discontinuous toward the north of the image, possibly due to increased erosion in this area. The M unit is sufficiently eroded in some locations to reveal an underlying light-toned layer (shown in Figs. 3–7).

Three types of terrain are visible at the northern margin of the smooth terrain, M (Fig. 3). The M unit appears to be composed of a series of small mesas truncated by erosion that shows the mesas to be composed of thinly stratified deposits (Fig. 3b). This is
consistent with the morphologic expression of mesas because this type of erosional landform commonly exists when the lithologies are subhorizontal, even if the layering is not visible everywhere. This unit appears to be ~50–100 m thick based on the topographic data. Many buttes (B in Fig. 3) exhibit a rougher, rockier texture than unit M. These buttes are generally <1 km wide and display no obvious internal layering. A light-toned layer with a strikingly higher albedo than the surrounding materials is present in between units M and B (white arrows in Fig. 3a and b). The outcrops of light-toned rocks are usually ~50 m in width (lateral extent) but are extensive enough such that they often surround buttes like an apparent aureole.

3.2. Stratigraphy of the light-toned layer

The presence of different light-toned outcrops around different buttes could be interpreted as the result of many light-toned layers devoid of relationships. However, in all locations the light-toned outcrops separate the loose deposits forming small mesas (M) from the rocky buttes (B). For this geometric reason, it is likely that the light-toned material represents a distinct stratigraphic layer. The total spatial extent of the exposed locations of this layer is about 20 km by 30 km. HiRISE images suggest a thickness of a few meters.

Because this light-toned material often occurs at the base and surrounds the lower portions of buttes, it could be interpreted to be stratigraphically beneath the buttes. However, we show hereafter that the geometry is more complex and that, in fact, the light-toned layer is located stratigraphically above the buttes (B) and below the layered deposits of the mesa (M). Fig. 4 shows the HRSC-derived topography in the northern edge of the study area (Fig. 2) where all of the units are well resolved. It displays a series of buttes that have a rocky texture and raised relief that is apparent in the topographic contours. The HRSC DEM is not accurate enough to distinguish each scarp of the HiRISE image, but it is accurate enough to give a general view of the topography. Each butte can be distinguished by the 50 m spaced contours, especially the highest butte at the bottom right. This topography shows a change in elevation of 300 m between the light-toned layer at the top of the image and the one visible at the bottom. It also shows that the light-toned layer cannot be a layer outcropping both on top of and at the base of buttes.

The close-up in Fig. 4a shows a place where the light-toned layer is located above the rocky scarp and again at the bottom of this same scarp, despite the ~50 m difference in elevation. A uniformly dipping stratum would not produce this geometry and instead the light-toned unit appears to be draped over the rocky scarp. Fig. 4b shows another close-up where the light-toned layer is topographically above the rocky hills. Here, a chevron shape shows that the dip of the light-toned layer is tilted to the west and overlies the rocky hill. Fig. 4c displays a section of the light-toned layer on the top rocky hills (shown by “low”) while some of the light-toned layer is located at the topographic base of the hill (as indicated by “high”). In most of these close-ups, the light-toned layer is superimposed by a unit that is smoother than the rocky buttes but which locally contains meter-sized blocks (as in Fig. 4c). This smooth unit with local layering is similar to the unit M in Fig. 3.

The contact between the light-toned material and the rocky buttes is even clearer in an area near the lava flow front. In this location, the light-toned unit is stratigraphically higher than the rocky buttes and is thus younger than the buttes (Fig. 5). The light-toned layer is partially covered by a darker medium-toned material. Fig. 5 also shows that the light-toned outcrop is disrupted by fractures at the meter scale. The overlying medium-toned material is not fractured suggesting it likely consists of looser material.

Fig. 6 shows three examples of buttes surrounded by the light-toned layer. In all of these examples, the light-toned layer is observed halfway up the butte slope and not at the base. In the right side of Fig. 6a, a piece of smooth layered deposits display a strong dip to the east. It is superimposed on the light-toned layer and the butte material. Fig. 6c shows a very similar geometry. A much more complex geometry is presented in Fig. 6b. Here, erosion has removed most of the overlying smooth deposits, creating a complex pattern of the light-toned material. However, it is clear from its geometry that the light-toned layer cannot be a layer outcropping beneath the butte. In these three examples, the light-toned layer appears as a material that is deposited over buttes, regardless of pre-existing topography. These observations and the relationships between the buttes and the light-toned layer are stratigraphically consistent with the interpretation shown in Fig. 6d: the bedrock with rocky buttes is draped by the light-toned unit, which in turn is draped by the dark, smooth layered deposits. Thus the light-toned layer occurs at the boundary between the underlying butte-forming material and the overlying stratified deposits.

Fig. 7 shows similar examples for smaller outcrops. The topographic relief of the bedrock is lower in these locations and erosion has stripped back the upper smooth material just enough to expose the underlying bedrock. These examples show that the light-toned layer is stratigraphically above the bedrock, as schematized in Fig. 7d.

Thus, the examples with the outcrops in Fig. 4, the flat hill in Fig. 5, and the draping morphology presented in Figs. 6 and 7 show that the light-toned layer is located stratigraphically above the
rocky buttes and beneath the smooth material. We interpret the rocky buttes to represent local bedrock, which has been draped by a smooth layered unit. This smooth material (with local meter scale blocks) consists of the unit M and erodes to form mesas towards the south (as seen from Figs. 3 and 7). The two interpretive sketches (Figs. 6d and 7d) show the geometric reasons why the light-toned layer occurs as an apparent aureole surrounding buttes: by being a layer at the base of the smooth layered deposits draping buttes, the light-toned layer always outcrops at the contact between these two units when erosion allows these layers to be visible. In the south, the rocky bedrock and the light-toned layers do not outcrop anymore, likely because the overlying unit M has not yet been eroded.

3.3. Mineralogy of the light-toned layer

CRISM data were analyzed in the region where the light-toned layer frequently crops out at the surface. Spectra from several pixels show the presence of a 1.9 \( \mu \text{m} \) \( \text{H}_2\text{O} \) band in this material. A mapping of the 1.9 \( \mu \text{m} \) band depth over the HiRISE and HRSC images shows a good match between the location of the light-toned layer and increases in the strength of the hydration band (Fig. 8). Fig. 8b shows that almost all outcrops of the light-toned layer display an absorption feature attributed to hydrous minerals. To reduce the noise in this parameter and to produce a more conservative map of 1.9 \( \mu \text{m} \) hydration, only band depths greater than 1% are displayed in Fig. 8c. Very few pixels with hydration...
Fig. 5. Close-up of HiRISE image PSP006824_1725 showing a light-toned layer with fractures on a flat rocky butte. A thin unit that appears relatively loose and medium-toned is superimposed on the light-toned material. See Fig. 2 for locations.

Fig. 6. Close-ups of HiRISE image 6969_1725. The white arrow indicates medium-toned slightly layered material superimposed over the light-toned layer and over the butte. The light-toned layer, which surrounds the buttes, appears to drape it, as seen from the different elevation at which it is observed in (a)–(c). (d) Interpretative sketch. See Fig. 2 for locations.
signatures are detected outside the extent of the light-toned layer on this map. Pixels with a yellow color have band depths of ≥ 1.5%, which are uniquely associated with the light-toned layer. A close-up on two buttes of the central area in Fig. 8 shows how strong hydration signatures are well correlated to the light-toned layer (Fig. 9).
the surface is 10–40 m, which is just large enough to enable CRISM to map this layer by 1 or 2 pixels in the highest resolution mode. In addition to the 1.9 μm absorption band, spectra of the light-toned material also exhibit a 1.4 μm band (due to OH and/or H2O) and a strong decrease in reflectance from ~2.3 to ~2.4 μm (Fig. 10). The three examples shown have similar overall shapes with similar bands. In general, all light-toned outcrops display similar spectra suggesting a common mineralogy, likely due to a common origin. This similarity is apparent when comparing individual spectra (as for A) and spectral averages (as for B and C). In these spectra, both the 1.4 and 1.9 μm hydration features are observed, as expected for hydrated minerals. The strong drop at ~2.4 μm is typical of most hydrated sulfates and is also observed in some hydrated chloride salts. In polyhydrated sulfates, the 1.4, 1.9, and 2.4 μm absorption bands result from the combination of OH∗ or H2O bending, stretching and rotational fundamentals, or S–O bending overtones (e.g., Cloutis et al., 2006; Crowley, 1991). The spectra lack a ~2.1 μm absorption characteristic of monohydrated sulfates such as kieserite (Cloutis et al., 2006). Spectral ratios between the three spectra (A–C) and spectrally ‘neutral’ areas (e.g., dusty areas) in the surrounding terrain were calculated to further enhance these diagnostic absorptions. In addition to the three previous bands, these ratios display a shallow ~1.8 μm band (especially for A) that is only slightly visible in the individual spectra, though interpretation of this band must be used with caution given the level of noise.

A comparison of the CRISM ratio with selected library spectra of hydrated sulfates (bassanite CaSO4·1/2H2O, polyhalite K6Ca2Mg(SO4)4·2H2O, epsomite MgSO4·7H2O, copiapite Fe2+Fe3+)4(SO4)6(OH)2·2H2O, and polyhalite KCl·2H2O, carnallite KMgCl3·6H2O) shows that various minerals display the observed absorption bands (Fig. 10). Nevertheless, epsomite and copiapite have a stronger 1.4 μm band than what is observed in spectra A, B and C. The polyhalite exhibits a strong drop in reflectance shortward of 2.4 μm. Bassanite appears to be a better match than the other hydrated sulfates, especially because it displays a small band at 1.77 μm as observed in our spectra. Gypsum has similar bands as bassanite but with additional absorptions at ~2.21–2.27 μm (e.g., Langevin et al., 2005) that is not present in spectra A–C. However, we note that such bands can be significantly weaker when gypsum is mixed with other phases and thus we cannot rule out a contribution of gypsum. Alternatively, several hydrated chlorides such as carnallite and antarcticite also have a sharp, narrow absorption at 1.76 μm (Crowley, 1991), but their overall shapes differ from the observed spectra. If the absorption near 1.8 μm in the CRISM data is a real feature and not spurious noise, then this light-toned layer contains hydrated minerals most consistent with Ca-sulfates such as bassanite or gypsum.

3.4. Age of the different units

The contacts and relative stratigraphy observed in the HiRISE images show the following units: (1) the unit with rocky buttes, (2) a light-toned layer at the base of, (3) the layered unit M, and (4) the volcanic plain, which is present only at the lowest elevations and is superimposed on all other units. These units must have formed after the Chasma opened because the light-toned layer and the underlying unit M are superimposed over the hillslope debris aprons. Absolute ages can then provide additional constraints on this chronology, especially the relationship between the layered deposits and the overlying volcanic plain.

The ages of the different units were obtained by counting craters 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. The determined ages are only valid for fresh surfaces that have not been affected by erosion or deposition. Crater counts should follow isochrons when the surface is fresh, whereas crater counts will cross isochrons when a surface has been subsequently modified.

The lava plain unit (VP) has been dated using HRSC and MOC images (Fig. 11) and corresponds to an isochron age between 50 and 100 Myr. The absolute age of the light-toned layer is more difficult to establish because it is part of the unit M, which has been substantially eroded. To minimize errors from counting craters on eroded surfaces, we have counted only the surface of the unit M where it was not visibly eroded, such as the flat mesa in Fig. 3a, and all mesas to the south of the Fig. 3a close-up. These areas might still have been slightly modified by erosion but they are comparatively less modified than those which are visibly eroded. Assuming that the light-toned layer is part of the M unit, and barring any temporally significant unconformities within this unit, the age given by crater counts should give the minimum age of the light-toned layer. The age of unit M obtained from the isochrons

![Fig. 9.](image)

(a) Close-ups over two buttes in the center of Fig. 8. (b) CRISM hydration band depths from 1% (orange) to 2% or more (yellow). Hydration signatures correspond to the light-toned layer. (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)
is between 10 and 100 Myr and is in the same range as the age of the VP unit (lava plain). Despite a small age difference might exist and that the similarities in age may be a coincidence, the stratigraphic relationships, spatial proximity, and similar ages of these two units suggest they may be genetically related.

4. Discussion

4.1. Processes of aqueous ash alteration

We propose that the light-toned layer and the rest of the overlying unit M were emplaced by airfall deposition of volcanic ash. Indeed, the center of activity of the volcanic plain is <10 km to the west of the light-toned unit. The estimated thickness of this deposit (tens of meters) is within the range of proximal ash deposits on Earth (e.g. Fisher and Schmincke, 1994). In such interpretation, the presence of occasional meter scale blocks inside these deposits (Fig. 4a and b) could correspond to volcanic bombs interstratified into ashes and expelled from the nearby volcanic sources. In addition, the occurrence of sulfates such as gypsum in terrestrial ash beds has been long recognized as a common alteration product (e.g. Bao et al., 2003). SO₂ and HCl gas are common volcanic gases and would be expected to alter ash to form sulfates or chlorides as efflorescent salts or volcanic sublimates. In addition, the light-toned layer is fractured and cohesive compared to the upper deposits, which appear looser and friable. It is also observed on Earth that loose ash beds become more consolidated as they are cemented and altered into sulfate-bearing rocks (e.g. Bao, 2005).

The deposits are located in the floor of a 6 km deep Chasma, therefore seepage from canyon walls or local ponds are possible water sources. However, the light-toned sulfate-bearing layer does not follow any constant height level that would suggest a water level of a standing body of water. The difference of elevation of 400 m between the highest and lowest two outcrops of the light-toned layer is not consistent with a seepage from a defined...
lithologic level. Despite paleolakes have locally been identified in Valles Marineris and Tharsis regions (e.g. Quantin et al., 2005; Mangold and Ansan, 2006), the lack of fluvial landforms associated to the Chasma (valleys, delta fans) does not suggest a similar context. No shoreline defines possible paleolake contour around the uppermost light-toned layers. Therefore, no classical evaporitic sequence with ponds or groundwater circulation with brines is possible from the observed geometry.

Hydrothermal activity can generate an alteration at the contact of hot water and volcanic gases. On Earth, a significant amount of sulfur is captured as native sulfur, sulfate minerals, and sulfides as gases and fluids move through glacier-capped volcanoes (e.g. Zimbelman et al., 2005). Gypsum and anhydrite have been described around fumaroles at numerous volcanoes (e.g. Goff and McMurtry, 2000). Sulfate-bearing hydrothermal alteration in active stratovolcanoes occurs episodically when sulfur-rich volcanic gases are introduced into meteoric water, with subsequent formation of acids that are neutralized by reaction with volcanic rock (Zimbelman et al., 2005). This type of alteration is possible in the case of the observed activity in Noctis Labyrinthus. Nevertheless, the geometry and spatial extent of a unique thin sulfate-bearing layer is not consistent with alteration by local fumaroles. The observed layer may have formed by the eolian reworking of such material, but this reworking process would not explain the fact that the light-toned layer consistently drapes the underlying bedrock. Direct post-depositional alteration of ash is more likely, but this process is not the same as classical hydrothermal alteration.

Ash beds deposited close to the center of volcanic emissions could be warm enough to melt water ice/snow if it was present at the surface, or, eventually, if snow precipitation occurred shortly after the ash deposition. Snow deposition is not usual in equatorial regions due to the low atmospheric water vapor pressure coupled with a high mean annual temperature (above frost point), leading to the immediate sublimation of ice at the surface in the current climatic regime. However, periods of high obliquity favor equatorial deposition of ice, forming local tropical glaciers (Forget et al., 2006). Recent models show the possibility that ice from relict glaciers identified in the regions of the giant volcanoes would have partially re-condensed as snow in other regions, including nearby Noctis Labyrinthus (Fig. 2 in Forget et al. (2007)). Though no large glaciers are visible in the Noctis region, one tongue-like feature interpreted as a relic glacial deposit is observed in the vicinity of the study region (Fig. 12). Indeed, its shape is different from the usual shape of landslides in Valles Marineris canyons (e.g. Quantin et al., 2004). For example, a fine layering is visible in that tongue, but never visible in any giant landslides, and might represent viscous flow features as observed at mid latitudes (Milliken et al., 2003; Mangold, 2003) that would require a proportion of water ice larger than 28% (Mangold et al., 2002). The steep end resembles the end of several tongues visible east of Hellas (Forget et al., 2006). Thus, the shape of this tongue, observed at the foot of a 4 km high plateau, is consistent with that of a glacier, and the timing of the volcanic activity in Noctis (between 50 and 100 Myr) could fit within potential periods of high obliquity on Mars (Levard et al., 2004). Interaction of local snow deposits with hot ash containing sulfur and chlorine might have created the environment required for the partial alteration of the ash bed.

Dry fog hypotheses have also been proposed for terrestrial sulfate deposits as well as for Mars (e.g., Banin et al., 1997; Bao et al., 2000). Sulfates in ash beds have sometimes been interpreted as due to modification of playas, but a more likely source is the wet and dry atmospheric deposition of sulfate produced by atmospheric oxidation of reduced gaseous sulfur compounds (Bao et al., 2003). A volcanic origin for some sulfate in terrestrial volcanic regions is supported by sulfur isotope data (Bao et al., 2000). The key components in the process are a surface area of fine ash particles and sulfate aerosols derived from the oxidation of volcanic SO2. Potential oxidation pathways for volcanic SO2 include oxidation by air O2 on catalyzed particle surfaces, aqueous oxidation by H2O2 or O3, and gas-phase oxidation by the OH radical (Bekki, 1995; Harrison and Larson, 1974; Liang and Jacobson, 1999). On Mars, the problem is that the ash deposits might not have reached high atmospheric elevations required for this process (Bao et al., 2003) and that the martian atmosphere may lack sufficient water vapor to generate this process as proposed on Earth. Nevertheless, recent geochemical models try to explain the formation of sulfates on Mars by solid–gas interactions (Berger et al., 2008). In this model, water is derived from the melting of ice, whereas SO2 is produced by the dry oxidation of H2S and SO2 in the atmosphere. In contact with near-surface water, the gas will dissolve extremely rapidly, producing pristine sulfuric acid capable of altering volcanic particles at the martian surface. Thus, this theory would involve sulfur-rich gases and snowmelt rather than atmospheric water, leading to similar conclusions as the previous hypothesis.

In summary, a traditional evaporite sequence is unlikely to have formed the observed alteration layer. The hypothesis that fits the observations best is alteration due to melting of local frost/snow and further water circulation, possibly helped by sulfur gas/rock interactions. The fact that the upper sequence of layers does not display hydrated alteration products favors this interpretation because the water ice would only be present at the beginning of the alteration process. Later episodes of ash deposition would not lead to alteration because of the source of water (snow/frost) would be absent (Fig. 13).

4.2. Implications for light-toned layers formation

Early investigations of the OMEGA dataset led to the identification of three different types of sulfates on Mars (Gendrin et al., 2005): monohydrated sulfates (where kieserite MgSO4·H2O, is the best spectral match for strong absorptions), gypsum (CaSO4·2H2O), and other polyhydrated sulfates which can correspond to Mg-, Na-, Fe-, or mixed-cation sulfates. These sulfates were detected in light-toned deposits and had large exposures in Terra Meridiana, the Chaos region, and in the Valles Marineris Chasmata. From a
geological point of view, no geological criteria permits distinction of the light-toned layer in Noctis Labyrinthus studied here from individual sulfate-bearing layers in the light-toned deposits elsewhere in Valles Marineris. Nevertheless, the main sulfate-bearing layered deposits in Valles Marineris such as those in West Candor Chasma (Mangold et al., 2008a; Le Deit et al., 2008) are >5 km thick and thus are many orders of magnitude thicker than the several meters thick layer described here. The spectral difference between the hydrated phase(s) identified in Noctis Labyrinthus and those identified elsewhere in Valles Marineris is slight, limited to the presence of the ~1.8 µm band, but it suggests a difference in aqueous geochemistry, precursor composition, or both. Indeed, spectra of the volcanic plains in the Noctis area are dominated by high calcium pyroxene (Mangold et al., 2009; Poulet et al., 2009). Alteration of such high calcium materials would be consistent with the presence of Ca-sulfates such as bassanite or gypsum, assuming the ash composition was similar to that of the volcanic plains unit.

OMEGA and CRISM mineralogical mapping has shown sulfate deposits are mostly restricted to the Late Noachian–Late Hesperian (e.g. Gendrin et al., 2005; Bibring et al., 2006). A mineralogical timeline has been proposed where this period, called the theiikan, represented a global environment with more acidic water (Bibring et al., 2006). The alteration deposit in Noctis Labyrinthus is much younger than the theiikan period; this example shows that sulfates and related hydrated mineral phases can form under local processes, without being related to a global environment. Such examples may not be common because current erosion rates of young deposits are slow and any buried alteration layer would be difficult to be observed at the surface. Nevertheless, the salts and other secondary alteration products found in several martian meteorites such as shergottites and nakhlites (e.g. Bridges et al., 2001) may have been produced by such type of volcanic/water interactions in the recent past.

This local formation of hydrated salts also raises the question: are sulfate deposits elsewhere in Valles Marineris indicative of the global climate/environment or are they the consequence of many local environments under a period of increased volcanic activity? Indeed, the studied example shows the possibility of alteration in a recent period where the climate was presumably as dry and cold as today. The thickness of the observed sulfate-bearing layer is small enough relative to the overall amount of volcanic material that an analogous process for more ancient sulfates would require a tremendous amount of volcanic material, as might have existed, but also a much larger amount of water than today. Nevertheless, this isolated example provides a tool to better understand processes that may have been more widespread in the past. On the other hand, gypsum-bearing dunes found in northern plains (Langevin et al., 2005) might be related to a similar process as observed in Noctis, with the difference that the volcanic source and volcanic lithologies are no longer visible in those locations.

5. Conclusion

A light-toned layer is identified in one Chasma of the Noctis Labyrinthus region. Stratigraphy and topography derived from HiRISE and HRSC data show that this light-toned layer is part of a thinly stratified unit that drapes pre-existing hilly bedrock. Spectral data show that the light-toned layer contains minerals whose spectral signatures are consistent with hydrated sulfates or chloride salts, though Ca-sulfates such as bassanite and gypsum provide the best matches. The vicinity of a smooth volcanic plain, similar ages as derived from crater counts, and stratigraphic relationships suggest the alteration layer may have been produced during the same period of volcanic activity that formed the nearby volcanic plain. In agreement with morphology, the light-toned layer and thinly stratified unit may represent ash deposits. Alteration of the lowermost layer, which is now the light-toned layer, may have occurred during interaction of water ice with hot ash, either from pre-existing tropical snow deposits, or snow precipitation shortly after ash deposition, or alternatively by solid–gas interactions due to nearby volcanic degassing. The alteration minerals in Noctis Labyrinthus may be different from those observed elsewhere in Valles Marineris because of local differences in volcanic chemistry, abundance and availability of water, and/or emplacement mechanism. The young age of the volcanic plain (Late Amazonian, <100 Myr) implies that recent alteration processes can exist on Mars in relation with recent volcanic activity. Such localized deposits do not require a martian climate that is different than today and may be facilitated by obliquity variations capable of mobilizing water ice to low latitudes.
Acknowledgments

We acknowledge the effort of the HRSC Co-Investigator Team members and their associates who have contributed to this investigation in the preparatory phase. We thank F. Scholten and R. Schmidt for their technical help and advice, in using DLR’s photogrammetry software, and two anonymous reviewers for their detailed comments. French authors were supported by the Programme National de Planétologie (PNP) of the Institut National des Sciences de l’Univers (INSU) and the Centre National d’Etude Spatial (CNES).

References


Mangiold, N., and 11 colleagues, 2008b. Local sulfate-rich layered deposits in Noctis Labyrinthus, Mars, and their chronological consequences. AGU Fall Meeting, #P44A-03.


