Properties of the Nili Fossae Olivine-phyllosilicate-carbonate lithology: orbital and in situ at Séítah

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Key Points:

- **We use CRISM data and a Venn diagram analysis to show the Nili Fossae olivine lithology is better named olivine-phyllosilicate-carbonate**
- **We postulate a flood lava origin for the olivine-phyllosilicate-carbonate at Séítah based on finding olivine cumulate with low viscosity**
- **We find the phyllosilicate in the cumulate olivine at Séítah is likely talc, serpentine, hectorite, Fe/Mg smectite, saponite or stevensite**

Plain Language Summary

We used orbital and in situ data to observe a lava flow containing the mineral olivine near the Mars 2020 landing site at Jezero Crater. By analyzing the reflectance spectra of the rocks containing the lava, we have identified that clay is present in the olivine bearing rocks. We use in situ imaging data to determine that the lava contains the olivine in close packed crystals (cumulate), a process which can happen in the bottom of a lake of lava. We use measurements from the SuperCam instrument to determine that the cumulate is accompanied by clays.

Abstract

We have studied the observed properties of the Nili Fossae olivine-phyllosilicate-carbonate lithology from orbital data and in situ by the Mars 2020 rover at the Séítah unit in Jezero crater, including: 1) composition 2) grain size 3) inferred viscosity (calculated based on geochemistry collected by SuperCam (Wiens et al., 2022)). Based on the low viscosity and distribution of the unit we postulate a flood lava origin for the olivine-phyllosilicate-carbonate at Séítah.

We include a new CRISM map of the phyllosilicate 2.38 μm band and use in situ data from Mars 2020 SuperCam Laser Induced Breakdown Spectroscopy (LIBS) and VISIR and MastCam-Z observations to show that the phyllosilicate in the olivine cumulate in the Séítah formation is either talc, serpentine, hectorite, Fe/Mg smectite, saponite or stevensite.

We discuss two intertwining aspects of the history of the lithology: 1) the emplacement and properties of the cumulate layer within a lava lake, based on terrestrial analogs in the Pilbara, Western Australia, and using previously published models of flood lavas and lava lakes, and 2) the limited extent of post emplacement alteration, including phyllosilicate and carbonate alteration.

1. Introduction

1.1 Purpose of paper

This paper uses the latest findings of the Mars 2020 rover mission to infer the properties of the Nili Fossae olivine-carbonate lithology, including composition (Liu et al., 2022 and Corpolongo et al., this issue) and grain size (Tice et al., 2022), and uses this to calculate magma viscosity, based on the geochemistry of Wiens et al., (2022). We also use orbital CRISM data alongside in situ SuperCam reflectance measurements to derive new knowledge regarding the accompanying phyllosilicates in the olivine-carbonate lithology.

Importance of Viscosity. As we shall discuss further in this paper, the viscosity is a key characteristic of a lava flow, determined by a variety of factors, including its composition, temperature, water, crystal content and pressure. The viscosity is a critical parameter for controlling the morphologic expression of the erupted unit. Critically, it allows us to compare the lava with other similar flows across Mars, and place it in context with the dominant rock-forming processes on the planet. The inferred viscosity has now become available from the first in situ measurements of Jezero crater floor geochemistry.

Olivine-phyllosilicate-carbonate lithology and relationship to Landing Site. The olivine-phyllosilicate-carbonate lithology was discovered using the Thermal Emission Spectrometer on the MGS spacecraft (Hoefen et al., 2003). It was subsequently mapped to extend throughout the Nili Fossae region, a region of radial graben on the northwest side of the Isidis Basin (Kremer et al., 2019). The landing site of the Mars 2020 rover was chosen to be at Jezero Crater, the site of a delta deposit which indicated the crater previously was filled with liquid water. The rover landed on Feb 18, 2021, sol 0 of the mission, to the south east of the delta (Sun et al., this issue). Around 200 sols (Martian days) later, the rover traversed around the edge of the Séítah region (Figure 1a), which is covered in linear dunes that were deemed to be too difficult to traverse. The dunes were determined from orbit to have a similar spectral signature to the olivine lithology mapped outside the crater (Godge et al., 2015).

Previously the lithology has been known to be olivine-bearing (Hoefen et al., 2003; Tornabene et al., 2008) or olivine-carbonate (Ehlmann et al., 2008; Mandon et al., 2020; Brown et al., 2020), however, this paper is designed to quantitatively analyze the relationship between olivine, carbonate and phyllosilicate and investigate, using a Venn diagrammatic approach; the suggestion that where there is carbonate, there is also clay (Horgan et al., 2020; Tarnas et al., 2021).

We address in detail two aspects of the olivine-phyllosilicate-carbonate lithologic history: 1.) the emplacement of the cumulate layer within a lava lake, which is part of a flood lava, based on terrestrial analogs in the Pilbara, Western Australia, and previously published models of flood lava eruption and flow, and lava ponding in lakes (Worster et al., 1993), and 2.) the limited extent of post emplacement alteration, including phyllosilicate and carbonate alteration of the Séítah formation (Clave et al., this issue; Mandon et al., this issue). These aspects of the history of the unit demonstrate that observations and models are both needed to uncover further insights into the emplacement and subsequent alteration history of this important Martian lithology, which has now been sampled for return to Earth (Simon et al., this issue).

Unit Interpretations. Farley et al. (2022) list definitions for the rock lithologies discussed in this paper. Farley et al. defined the Séítah (an olivine cumulate) and Máaz (a pyroxene- and plagioclase-dominated unit overlying Séítah) formations that were encountered and mapped during the Crater Floor campaign in the first 380 sols of the Mars 2020 mission. We discuss interpretations arising from the Séítah formation, which has previously been interpreted as related to the olivine-carbonate lithology inside and outside Jezero. Figure 1 presents a visual summary of the traverse through the Séítah formation and the three key workspaces, and all the targets referred to in this paper. On Sol 202 of the mission the rover rolled into the Séítah formation, and approached the Bastide workspace. For the purposes of this study, the rover visited three main workspaces - in temporal order, Bastide, Brac and Issole. These workspaces and the abrasions and samples taken at each workspace are also shown in Figure 1. The rover left Séítah on Sol 340. See also (Sun et al., this issue; Crumpler et al., this issue; Núñez et al., this issue) for more details on the Crater Floor campaign and stratigraphic information obtained during the Séítah traverse.

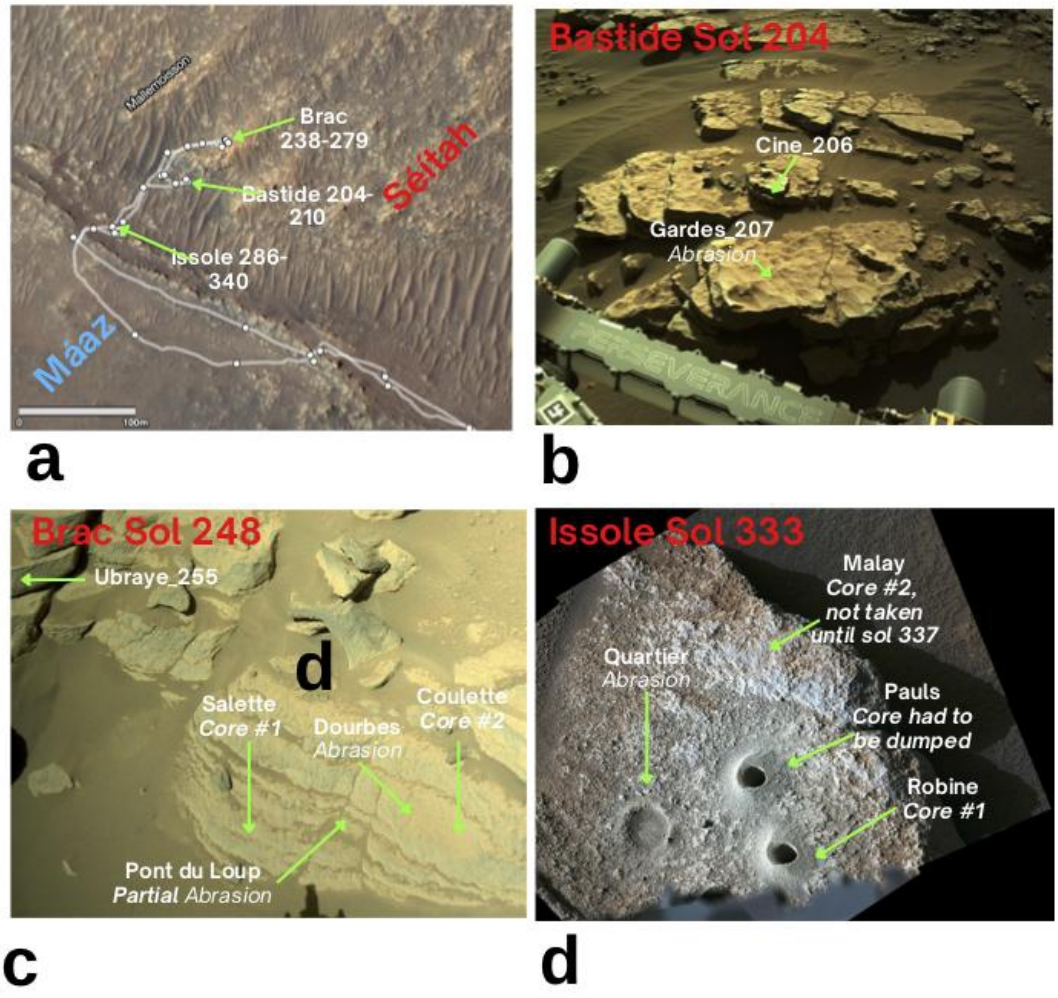


Figure 1. a.) HiRISE overview map of the Séítah traverse showing the three key workspaces: Bastide, Brac, and Issole, and the Sol counts at each. b.) Navcam image of Bastide Workspace taken on Sol 204. c.) Navcam image of Brac Workspace on Sol 248, before samples were taken d.) Mastcam-Z image of Issole on Sol 333 after Robine and Pauls samples and Quartier abrasion were taken but before sample Malay was obtained.

Definition of phyllosilicate. Due to its importance in this paper, we first establish a clear and distinct definition of the word "phyllosilicate". Here by "phyllosilicate" we are referring to the mineral group of phyllo-silicates ("φύλλο" in Greek is "leaf" in English). These layer-silicates have stacked hydroxyl (-OH), octahedral (O) sheets and tetrahedral (T) silicate (Bailey, 1988). The octahedral sheets come in two forms called dioctahedral and trioctahedral which host divalent (Fe^{2+}/Mg^{2+}) and trivalent (e.g., Al^{3+}/Fe^{3+}) cations respectively. Phyllosilicates come in many families, based on the stacking of layers (e.g. TO or 1:1 vs TOTTOT or 2:1). In this paper we are particularly interested in the 2:1 phyllosilicates talc and 1:1 serpentine and the 2:1 phyllosilicate family called smectites.

1.2 Previous work

1.2.1 Infrared remote observations

The Syrtis Major shield region was first observed to host olivine using Visible and Near InfraRed (VNIR) telescopic observations by Pinet and Chevrel (1990). Spacecraft observations followed, and a decade later the Thermal Emission Spectrometer (TES) infrared (IR) instrument also recognised a strong 10 μm absorption band associated with the Nili Fossae region that was attributed to olivine (Hoefen et al., 2003). The Observatoire pour la Minéralogie, l'Eau, le Glace et l' Activité (OMEGA) instrument on Mars Express was able to determine that phyllosilicates were also associated with the olivine, suggesting Fe-rich smectites were present (Mangold et al., 2007). Using data from the Compact Reconnaissance Imaging Spectrometer for Mars (CRISM) instrument, Mg-carbonate was discovered to be associated with the olivine lithology by using the strength and position of the 2.5 μm band (Ehlmann et al., 2008). The olivine-carbonate was hypothesized to be due to serpentinization by Brown et al. (2010), who also suggested that talc might be present by analogy to talc-carbonate terrestrial Archean greenstone terranes. Brown et al. (2010) also found the carbonate to be a 50/50 Fe/Mg mixture, which was later supported using TES orbital data analyzed by Ruff et al. (2022). Goudge et al. (2015) mapped the watershed of the Jezero deltas and the accompanying mineralogy. Mafic minerals and marginal carbonates were identified in the mapping of Horgan et al. (2020). Mandon et al. (2020) used OMEGA observations to map the olivine and carbonate, and used crater counting to date the unit at 3.82 Ga. The olivine composition (Fo#44-66) and grain size (~ 1mm) were calculated by Brown et al. (2020) using CRISM data.

A recent map of the mineralogy of Jezero crater using CRISM data illustrates the locations of multiple minerals and mineral combinations including olivine, Mg-rich carbonate, olivine+Fe/Mg-smectite, high-Ca pyroxene, and low-Ca pyroxene (Parente et al., 2021). This map was created using a new algorithm for atmospheric correction (Itoh and Parente, 2021) that enhances the spectral features due to surface outcrops and an improved mapping procedure (Saranthan and Parente, 2021).

Finally, a recent closely related study of VNIR spectra of phyllosilicates in the Archean Barberton Greenstone Belt has identified serpentine and chlorite after olivine and suggested that the phyllosilicates were formed by hydrothermal alteration (Grosch et al., 2021).

1.2.2 SuperCam LIBS observations

SuperCam is a multi-technique instrument suite mounted on the Mars 2020 *Perseverance* rover (Wiens et al., 2021; Maurice et al., 2021). It is an upgraded version of the ChemCam instrument onboard the Mars Science Laboratory (MSL) *Curiosity* rover. It has the capability to obtain LIBS measurements to a distance of ~10 m, and to quantify the major element abundances at distances up to 6.5 m. The Remote Micro-Imager (RMI) is a color telescope used to provide context images for SuperCam observations.

The SuperCam instrument has been used to investigate the igneous nature of the Séítah and Máaz regions, as reported by Wiens et al. (2022). That study concentrated on chemistry and mineralogy, and creation of a stratigraphy of the two regions of the Jezero crater floor. In this paper we use LIBS elemental abundances for the Séítah and Máaz formations reported in the Wiens et al. (2022) study.

1.2.3 SuperCam VISIR observations

SuperCam VISIR is an optical spectrometer that covers the 0.4-0.85 and 1.3-2.6 μm wavelength ranges, providing reflectance spectra boresighted with LIBS and RMI instruments (Fouchet et al., 2022). This paper relies on the calibration and comparison of the spectra presented here with others measured on the surface. Royer et al. (this issue) shows that, based on band depth calculations and noise estimates, that most of the 2.46 μm detections we will discuss herein range from Signal to Noise Ratio (SNR) = 4 to SNR = 12. This spectral range is harder to study because of the fact that many spectral signatures are sampled with the minimum sampling (at most 2 spectral channels in the band) and several artifacts may be similar to absorption features (e.g., non-corrected spikes, calibration residuals). In our case, regarding the 2.46 μm region, we have SNR > 5 and as demonstrated below, detections of small bands in this region are realistic, given the accuracy of the calibration and the potential residuals. Positive detections must nevertheless be confirmed visually and their attribution validated by the presence of other features characteristic of phyllosilicates.

SuperCam's VISIR data allow on-ground comparison with absorption features seen from orbit by the CRISM instrument. Due to the fact that CRISM covers a larger spectral range (0.4-3.96 μm), in this paper we adopt a band fitting strategy over the 1.3 to 2.5 μm range.

1.2.4 PIXL observations

PIXL is a high resolution X-Ray fluorescence microscope in situ scanning instrument mounted on the arm turret of the rover (Allwood et al., 2020). PIXL observations of the Dourbes target have been used to identify the cumulate nature of the target and infer an igneous origin (Liu et al., 2022). The olivine composition is consistently homogeneous, and of Fo55+/-1 within a several mm analysis patch (Liu et al., 2022), which is a remarkable match with previous orbital work constraining the olivine Fo# to 44-66 (Brown et al., 2020). PIXL's X-ray diffraction capability, which can identify coherent crystalline domains, has been used to identify coarsely grained (1-2 mm), interlocking olivine grains (Tice et al., 2022), and SuperCam's RMI instrument measured a mean grain size of 1.45+/-0.20 mm over the rover's Seitah formation traverse (Wiens et al. 2022). These in situ observations are both good matches for the ~1 mm Nili Fossae olivine grain size calculation from orbital studies prior to landing (Brown et al., 2020).

2. Methods

2.1 Viscosity calculations

2.1.1 Bottinga and Weill empirical melt viscosity model

In order to calculate the inferred melt viscosity, we have used the empirical approach of Bottinga and Weill (1972), which calculates a viscosity vs. temperature curve for the magmatic silicate liquid. Their method uses five look-up tables according to the silica content of the magma. They then use these tables to sum the contributions of the elements common to terrestrial magma bodies to arrive at a inferred viscosity for a magma of that composition.

We have used the SuperCam LIBS geochemical elemental abundances for the Máaz formation, Artuby member and Séítah formation from Wiens et al. (2022). Wiens et al. (2022) made a partial correction for the LIBS compositions by removing all data points that were float rocks, soils, and apparent coatings, as well as all points that had major-element totals < 80 wt%. The latter step removes a large number of points with alteration, especially anything that included significant sulfate or carbonate contributions. Wiens et al. (2022) investigated how the removal of points affected the bulk compositions, and determined that the 80 wt% was a reasonable limit. In this study, we have made no further attempt to petrologically correct the measured bulk compositions to the original magma, and

will instead pursue this approach in future work.

2.1.2 Flood lava and lava lake models

Martian gravity is ~37.5 % of that on Earth. Assuming Mars possessed a thinner (~6 mbar) atmosphere throughout its history, we can compare terrestrial lava properties to those of Mars (Wilson and Head, 1994). On Mars, with a thinner atmosphere, lava will take longer to cool. Longer cooling rates should also drive larger crystal growth. Wilson and Head (1994) predicted Martian volcanic flows (driven by lower gravity) to be 6 times as long as compositionally identical flows on Earth, leading to spatially larger deposits on the smaller planet.

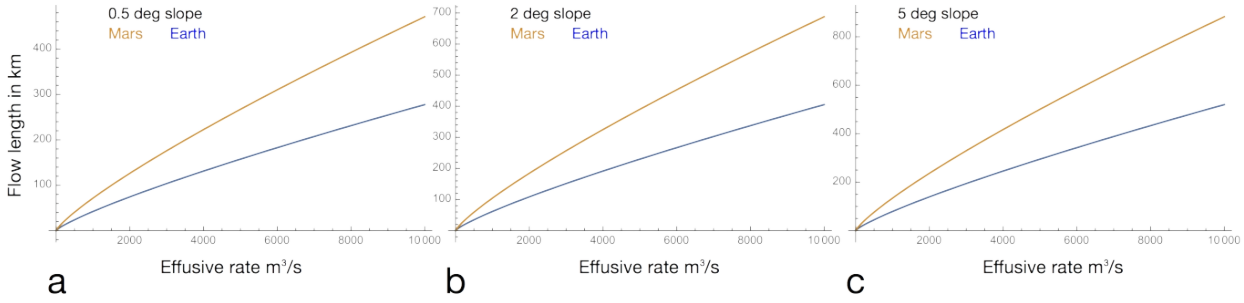


Figure 2. Lava flow lengths for Martian and Earth conditions for three different ground slopes. Modeled after Wilson and Head (1994), and recalculated using eqn (1). Angles are in degrees.

Pinkerton and Wilson (1994) presented a Bingham non-isothermal model for calculation of the runout lengths for lavas on Earth. This model depends upon viscosity, so we used this equation to calculate the expected runout length L in Figure 2 for Mars and Earth gravity. This is Equation 7 of Wilson and Head (1994):

$$L = [1.34/\kappa G_{zc}](\tau/\eta)^{2/11}E^{9/11}[\tau/(\rho g)]^{6/11}\sin^{3/11}\alpha \qquad (1)$$

where κ is the thermal diffusivity of lava, taken as 7×10^{-7} m^2/s, G_{zc} is the critical Grätz number, taken as 300, τ is the yield strength of the lava, taken as 2000 Pa, η is the plastic viscosity, $(\tau/\eta)^{2/11}$ is taken as 0.38 as suggested by Wilson and Head (1994), ρ is the density of the lava, taken here as a value appropriate for ultramafic lavas of 3200 kg/m^3, g is the acceleration due to gravity, 9.81m/s^2 for Earth and 3.72 m/s^2 for Mars, α is the average slope angle in degrees of the ground over which the lava flows. The results of this calculation are shown in Figure 2. Note that our values are a factor of 10 less than Figure 8b of Wilson and Head (which suggests flow lengths of up to 7000km, however only discusses flows up to 700km in the text of the paper) These show that run out flows of 300 to 700 km are

possible with the highest effusion rates, and that Martian flows are roughly 1.7 times as long as their terrestrial duals under the Pinkerton-Wilson model. In addition to this, a 5 times higher effusion rates is expected on Mars due to two fold greater dike widths expected due to lower gravity. These two factors lead to the factor of 6 times flow lengths expected on Mars. Martian lavas of length around 600km (such as Syrtis Major) would be equivalent to a 100km diameter lava on Earth (similar to the exposed current extent of the Warrawoona Group basalts in the Pilbara region (Van Kranendonk et al. 2002)).

In addition, the Pinkerton and Wilson model actually uses an implicitly high viscosity due to the $(\tau/\eta)^{2/11} = 0.38$ and $\tau = 2000$ we get $\eta \sim 409468$. For lower plastic viscosities we derive in this paper, the model predicts greater run out lengths are possible.

Finally, it should be noted that the Pinkerton and Wilson (1994) model is appropriate only in the case of "cooling-limited" flows, where the lava flow halts as it cools (Guest et al., 1987). If the lava flow is halted due to a magma cut off, the flow will be "volume-limited" and much shorter than calculated using equation (1).

The Reynolds number of a liquid flow can be calculated as (Huppert et al., 1984):

$$R_e = E/\nu \qquad (2)$$

Here E is the discharge rate and v is the kinematic viscosity. R_e is a threshold for turbulence, the critical value is typically ~2000.

At high Reynolds number, the discharge rate E of lava through a long fissure of width d can be calculated thus:

$$E = d^{3/2}(g\Delta\rho)^{1/2}/(k\rho)^{1/2} \qquad (3)$$

Here k is a friction coefficient, $\Delta\rho$ is the density difference between the magma and lithosphere, and ρ is the magma density. On Mars, as mentioned earlier, wider fissures d are expected by a factor of two, and this outweighs weak (square root) dependence on gravity, for otherwise identical flows, discharge rates will be 6 times than on Earth.

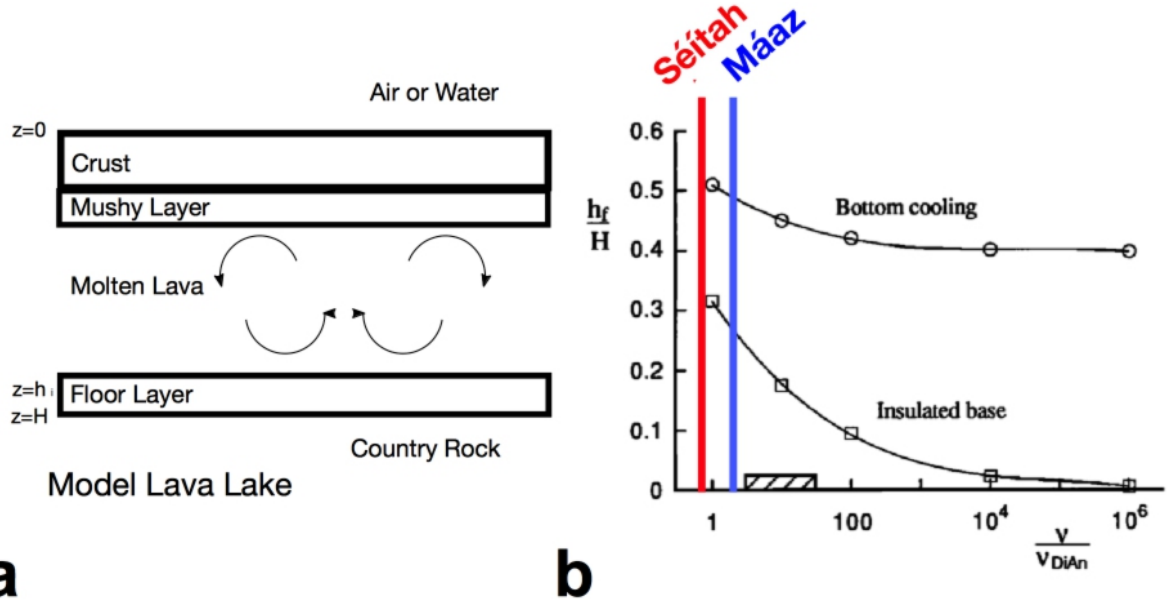


Figure 3. a.) Thermal model of a lava lake with layers representing solid crust/cap, mushy layer (where crystals form), molten lava (with convection taking place) and the floor. b.) Final fractional height of floor layer, h_f, where crystals accumulate (relative to the height of the chamber height, H) plotted against the lava viscosity (normalized relative to liquid diopside). Two cases are shown, one in which bottom cooling occurs, and one in which it does not (insulated base). The dashed box indicates the viscosity of typical terrestrial basalt relative to a pure diopside melt, after Worster et al. (1993). The Séítah and Máaz viscosities calculated in this paper are schematically shown on the plot.

Development of layering. The large grain size (Tice et al., 2022) and cumulate texture (Liu et al., 2022; Tice et al., 2022) lead us to explore the possibility of the Séítah rocks being emplaced as a Martian lava lake. Worster et al. (1993) developed a thermodynamic model for a terrestrial lava lake, and outlined how the body would form compositional and textural layers within the resultant rock as it cooled. They showed that a lava lake can develop quickly when a magma body is exposed to the atmosphere, and showed that convective cooling can take place. They examined the situation with and without convection in their model and found thermal convection drives internal crystallization within the melt. This in turn forms layering within the floor of the lake. Cooling by thermal conduction alone does not produce this layering. Figure 3a shows a schematic diagram of an idealized lava lake after Worster et al. (1993). For our purposes, we will define a lava lake as a connected igneous body that is cooling convectively and develops internal stratification (i.e. layering) as shown in this diagram.

Effect of viscosity. Figure 3b shows the effect of viscosity on the cooling history of

a lava lake that is cooled from the top. This figure shows the major effect of viscosity on lava lake crystallization, as studied by Worster et al. (1993). The parameter h_f is the final floor layer (Figure 3a) depth, and H is the total depth of the lava lake. It can be seen that the relative size of the floor layer grows with decreasing viscosity, and this effect is more pronounced when the bottom of the rock is insulated. We show schematically with vertical lines where the Séítah and Máaz units viscosity would plot (relative to liquid diopside), indicating that Séítah-like viscosities would produce a deeper cumulate floor layer than a Máaz-like viscosity. The equation for the calculation of the viscosity of diopside adopted by Worster et al. (1993) (their Table 2) is:

$$\nu = \exp[12\text{-}52.5x+62.5x^2]\ln(10)] \quad (4)$$

where $x = 1\text{-}1000/(T+273)$ and
T is the temperature of the melt in degrees C.

The takeaway message from this section is that lower viscosity lavas build thicker cumulate layers. We shall use this model to enhance the interpretations of our in situ observations later in this study.

2.2 Phyllosilicate alteration mineralogy

We will now discuss the methods used in this paper to analyse visible to near infrared (VNIR, 0.4-2.5 μm) reflectance spectra to identify and map phyllosilicates using asymmetric band fitting, and give a short introduction to the chemical formulas of the phyllosilicates that we are interested in for this study.

For this study we have used a single image cube of CRISM data from the Map Projected Targeted Data Record (MRTDR) calibration standard, as described by Seelos et al. (2012). This processing level removes all atmospheric effects and presents each pixel in apparent reflectance space, which is suitable for comparison with spectral library spectra or SuperCam VISIR processed spectra.

2.2.1 CRISM and VISIR phyllosilicate identification

Figure 4 shows library spectra of talc, lizardite, stevensite, Fe/Mg smectite, hectorite and saponite in the VIS and SWIR (Short Wave InfraRed) regions, which are covered by the SuperCam VISIR instrument. The talc and lizardite spectra are

sourced from the USGS Spectral Library (Clark et al 2007). The stevensite specimen was provided by Nicholas Tosca and measured at the University of Winnipeg Planetary Spectrophotometer Facility. The Fe/Mg smectite, hectorite and saponite spectra were supplied by Janice Bishop. Information on the samples is supplied in the supplementary section S2.

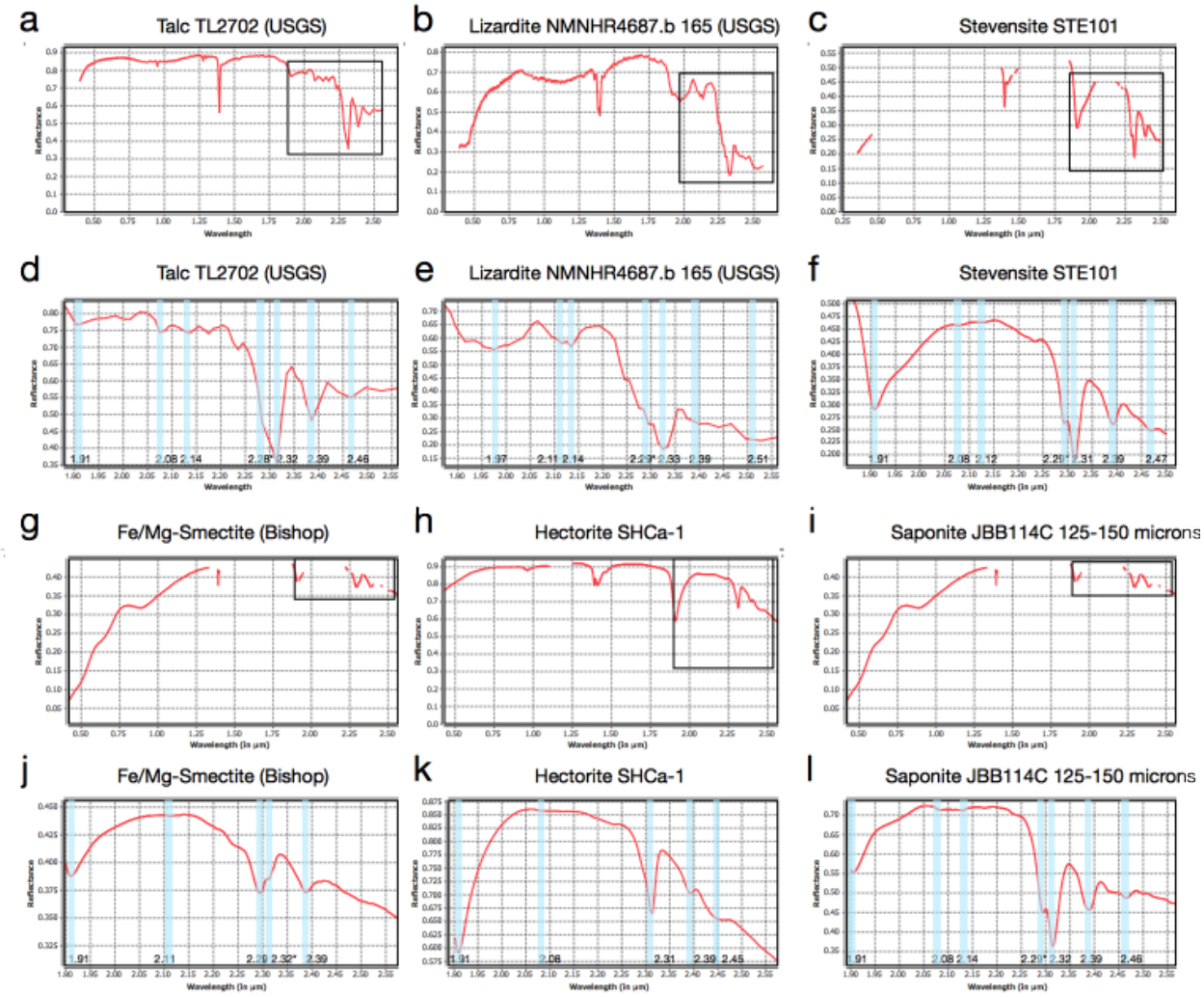


Figure 4. Laboratory spectra of a.,d.) Talc b.,e.) Lizardite c.,f.) Stevensite g.,j.) Fe/Mg-smectite (Bishop) h,k.) Hectorite SHCa-1, i,l.) Saponite JBB114C. Shown at top is the visible and SWIR region of the spectrum, on the bottom is a zoom into the SWIR region at the location shown in the box on the lower spectra. Note: 1) strength of 2.31 µm band in talc and 2) the presence of a weak 2.46 µm band in talc which is not present in lizardite.

The SWIR absorption bands of the Fe-bearing phyllosilicates nontronite (Bishop et al., 2013) and hisingerite (Turrenne et al., this issue) both display strong 2.28 µm bands which are not present in the spectra of olivine bearing targets we report here,

so they are not considered further.

The key bands of the spectra are noted in the plot, and we draw the reader's attention to the relative strength of the 2.31 (strong), 2.38 (moderate to weak) and 2.46 μm (weak) bands in particular. The two key differences for the purpose of this paper are 1) the strength of the 2.38 μm band, which is present in the serpentine lizardite (Figure 4e), but far stronger in talc (Figure 4d), and 2) the presence of a weak 2.46 μm band in talc (Figure 4e) which is only a shoulder (at most) in lizardite (Figure 4e). These spectral differences will be discussed in the remainder of the paper.

2.2.1.2 Band fitting with an Asymmetric Gaussian shape

As has been previously discussed in a variety of forums such as Brown (2006), the natural bandshape for an absorption band in energy space is a Gaussian shape. When accompanied by nearby and often closely related bands, for example, due to Tschermak substitution of Mg^{2+} and Fe^{2+} (Duke, 1994), the vibration bands can adopt an asymmetric shape. It is then advantageous to carry out an asymmetric band fit. Figure 4 demonstrates that all three of the phyllosilicate related bands are asymmetric and extend to the left (higher energy). As part of this fitting process we obtain four parameters per band: the amplitude, width, centroid (position), and the asymmetry (Brown et al., 2010). We primarily use the amplitude band parameter, and cross-correlate the amplitude of the olivine 1 μm band, 2.31 μm (phyllosilicate or carbonates), 2.5 μm (carbonate only), 2.38 μm (phyllosilicate only) band amplitude maps.

2.2.2 Ideal Chemical formulas of serpentine, talc and smectites

*Smectites***.** Dioctahedral smectites contain two cations with a charge of 3+ (typically Al^{3+} or Fe^{3+}) in their octahedral sheets (for example montmorillonite and nontronite), while trioctahedral smectites contain three cations with a charge of 2+ (typically Mg^{2+} and sometimes Fe^{2+} or other transition metals) with the Mg-rich form called hectorite, stevensite, or saponite, where saponite has tetrahedral substitution and hectorite does not. Stevensite has additional charge due to cation vacancies in the octahedral sheets.

Beidelite and saponite are special cases of smectites where Al^{3+} substitutes for Si^{4+} in the tetrahedral sheets, causing a disruption of the structure and creating surface charge. Nontronite and hectorite do not typically contain any Al^{3+} cations. We note

that the Fe/Mg-smectite proposed to best match the spectral properties of the most common phyllosilicate on Mars is a smectite containing a mixture of both Fe^{2+} and Mg^{2+} cations, but no Al^{3+} cations, termed "Fe/Mg-smectite" (Bishop et al., 2008a,b).

Saponite is also frequently considered the subgroup name for trioctahedral smectites; under this definition there are "saponites" with no Al^{3+} cations that resemble hectorite and there are high Al-substitution saponites that have significant substitution of Al^{3+} for Si^{4+} in the tetrahedral sheets. As it is difficult to determine the degree of Al-substitution in saponites without detailed sample characterization, this mineral name can be confusing. This confusion is often compounded because saponite is much more common on Earth than hectorite, and thus lab samples of saponite are more readily available for study. However, Mg-rich smectites do not require Al^{3+} in their structures, for example, saponite has tetrahedral substitution whereas hectorite does not.

Non-smectites. The ideal chemical formula of lizardite (Mg-serpentine) is $Mg_3Si_2O_5(OH)_4$, and that of Mg-talc is $Mg_3Si_4O_{10}(OH)_2$. For the purposes of this study, we point out that there is no aluminum in their structure. In contrast, the Mg-chlorite, clinochlore has an ideal formula of $(Mg,Fe^{2+})_5Al(Si_3Al)O_{10}(OH)_8$. A similar amount of Al^{3+} is present in micas such as the dioctahedral muscovite $KAl_2(AlSi_3O_{10})(F,OH)_2$ and trioctahedral biotite $K(Mg,Fe)_3AlSi_3O_{10}(F,OH)_2$.

3. Results

3.1 Viscosity results and lava lake calculations

Table 1 presents the elemental abundances derived using SuperCam LIBS measurements on rocks from the Máaz formation, Artuby member and Séítah formation. The LIBS data are obtained from Table 2 of Wiens et al. (2022). The elemental compositions are averages over many targets within the respective units. The CIPW norm mineral sum is slightly below 100 wt% because Wiens et al. assumed a small amount of apatite based on observations of phosphorous – Ca – F/Cl/H LIBS peak associations in a few targets; however, the P, Cl, F, and H abundances are not quantified, which leads to a small mass deficit.

The CIPW norms in Table 1 provide estimated igneous mineral content in the absence of alteration by carbonates or phyllosilicates. Wiens et al. (2022) removed observations that had clear evidence of alteration. In this paper, we will use this elemental abundance data to calculate the viscosity, and later we will discuss the

low Al content of the olivine bearing unit in the Séítah unit in relation to phyllosilicate identification.

The bottom line of the table shows the calculated viscosity at 1450 K. This calculation was conducted using the Bottinga-Weill empirical melt viscosity technique described earlier. The results of this calculation are plotted against the inverse temperature in Figure 5a. We have presented the viscosity results in this manner ("Arrhenius Plot" of natural log of viscosity versus the inverse temperature) because this is the format of the viscosity plots of Bottinga and Weill (for example their Figure 7). We then used equation (1) to calculate the length of the flow in the same way Figure 2 was calculated. We have used three different viscosities and two different gravities, and plotted the six lines in Figure 5b.

Relative viscosity analysis. As discussed in Section 2, viscosity has a large effect on the properties of lava lakes. We showed in Figure 3b the schematic relative viscosity values for the Máaz and Séítah formations relative to the viscosity of diopside. We are now in a position to justify this calculation. Using equation (4), we can calculate the model viscosity for diopside at 1450K as 382.54, and using the last line of Table 1, see that it lies between the viscosities at the same temperature of Séítah (88.55) and Máaz (472.68). We note that we are obliged in this comparison to estimate a temperature relevant for the lava lake model.

Viscosity Error analysis. In order to provide some measure of the error of these viscosity derivations, we use the error estimate from Anderson et al. (2022). The average Root Mean Squared Error of Prediction (RMSEP) from their Table 5 is 1.85 wt %. We also use the error estimate from Bottinga and Weill (1972) for their viscosity model. Figure 6 of Bottinga and Weill shows that 99% of their observations are within $-0.75 < \Delta \ln \eta < 0.75$ of their viscosity model. We therefore use +/- 0.75 as the error band for the ln viscosity

Table 1. Elemental abundances and normative mineralogies obtained using SuperCam LIBS during the Crater Floor campaign (Wiens et al., 2022). All elemental data are in wt % and the CIPW norm abundances are in %.

Element oxide (wt %)	Máaz	Artuby	Séítah
SiO_2	54.2	49.7	44.8
TiO_2	0.6	0.9	0.2
Al_2O_3	10.5	7.4	3.9
FeO_T	20.2	25.3	22.6
MgO	2.7	3.8	21.4

CaO	5.5	7.7	3.6
Na_2O	3.3	2.7	1.3
K_2O	1.2	0.6	0.2
Quartz	5.6	0.0	0.0
Plagioclase	37.9	30.0	15.2
Orthoclase	7.1	3.6	1.2
Diopside	4.8	16.7	1.5
Hypersthene	37.9	41.5	34.7
Olivine	0.0	0.9	41.1
Ilmenite	1.1	1.7	0.4
Magnetite	1.6	2.0	1.8
An # plag	25.3	22.7	26.5
Mg #	19.3	21.1	62.8
Density (g/cc)	3.10	3.27	3.37
	±0.02	±0.03	±0.01
Viscosity at 1450° K	472.68	18.37	88.55

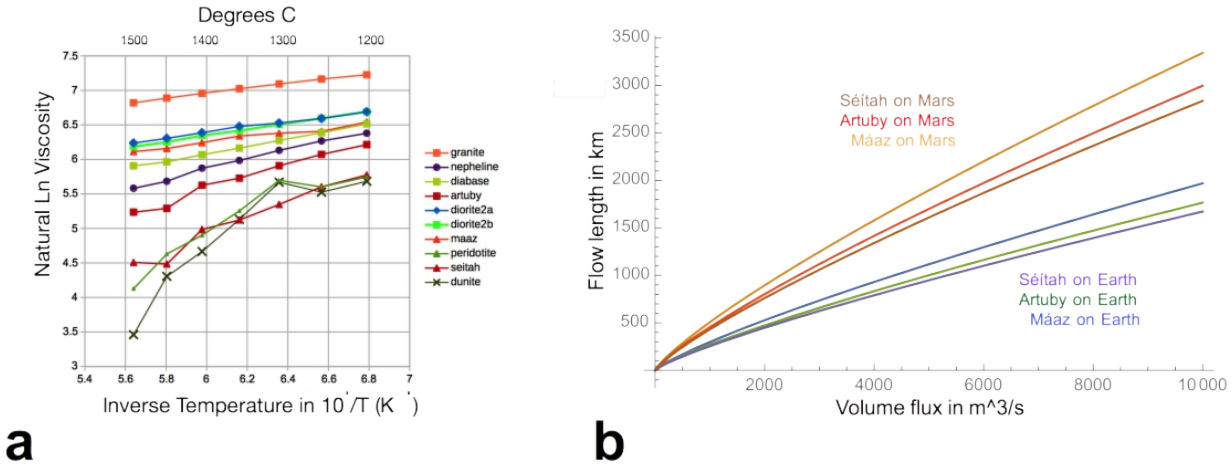


Figure 5. a.) Arrhenius plot of natural log of viscosity versus inverse temperature calculations for Máaz, Artuby, and Séítah rocks compared with terrestrial samples, including granite, diorite and dunite. Viscosity error bars are +/- 0.75 and are the same for all points. b.) Flow lengths for viscosities of Séítah, Artuby and Máaz on Earth and Mars.

3.2. Phyllosilicate alteration

3.2.1 CRISM orbital maps

To visualize the phyllosilicates that are present in the olivine-phyllosilicate-carbonate, we have produced a map of the orbital CRISM Half Resolution Long (HRL) image, HRL40FF, over Jezero crater (Figure 6). This includes the area of Mars 2020 operations, Octavia E. Butler Landing site, Jezero Delta, and the marginal carbonates region (Williford et al., 2018; Farley et al., 2020; Stack et al., 2020). Alongside a 0.905 μm channel image, we have included three absorption band maps from the CRISM image for 2.31 μm (phyllosilicate and carbonate), 2.5 μm (carbonate) and 2.38 μm (phyllosilicate). Here we focus on the phyllosilicate signature; for more discussion of the carbonate spectral signature, see (Mandon et al., 2020; Brown et al., 2020; Zastrow and Glotch, 2021).

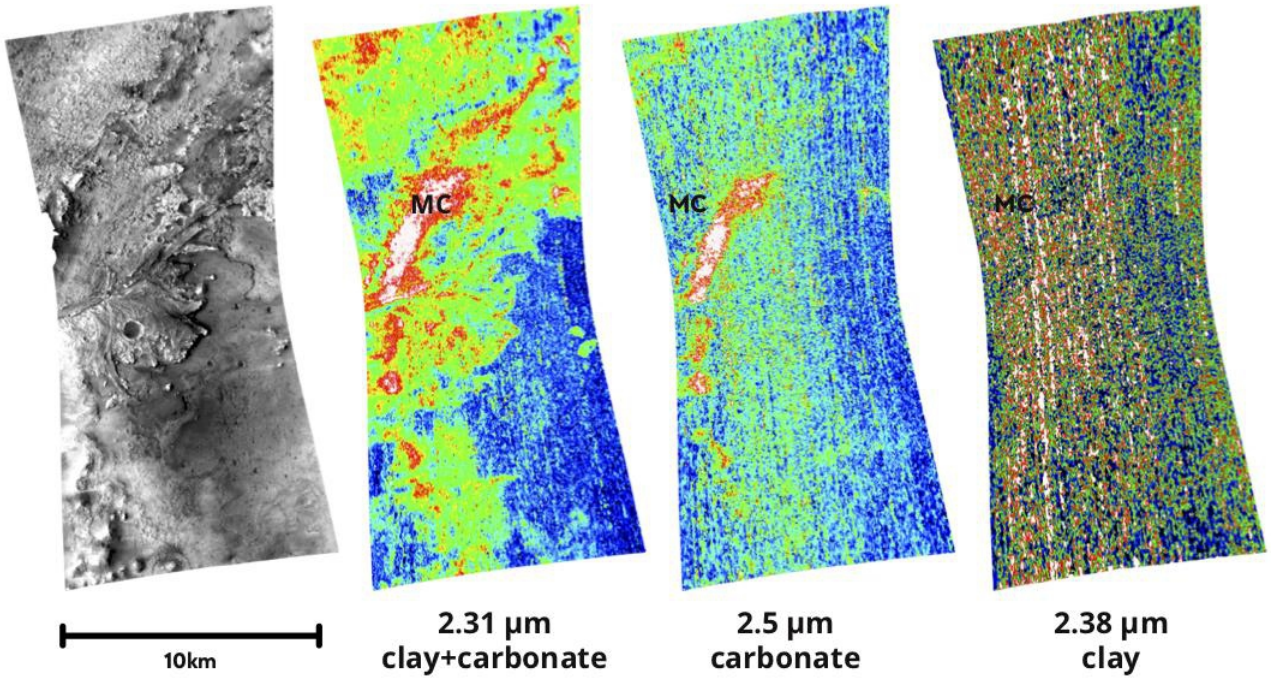


Figure 6. HRL40FF 0.905 μm map, 2.31 μm, 2.5 μm, 2.38 μm band amplitude maps. The location of the marginal carbonates is also shown as MC.

3.2.2 Correlation maps

As can be seen from Figure 6, mapping the 2.38 μm phyllosilicate band is very difficult due to it being close to the noise level. To overcome this and gain further

insights into the presence of phyllosilicates in the CRISM HRL40FF image, we have constructed a correlation map consisting of the 2.31, 2.38 and 2.5 μm bands. For each of these three band maps, we have the four parameters of the Asymmetric Gaussian band, meaning we have 12 parameters total. Figure 7 displays the correlation plots for HRL40FF for band depth amplitude (2.38)-amplitude(2.31)-amplitude(2.5). This can be interpreted as the *x* value being the amplitude of the 2.38 μm band, the *y* value as the amplitude of the 2.31 μm band, and the coloring of the points reflects the amplitude of the 2.5 μm band.

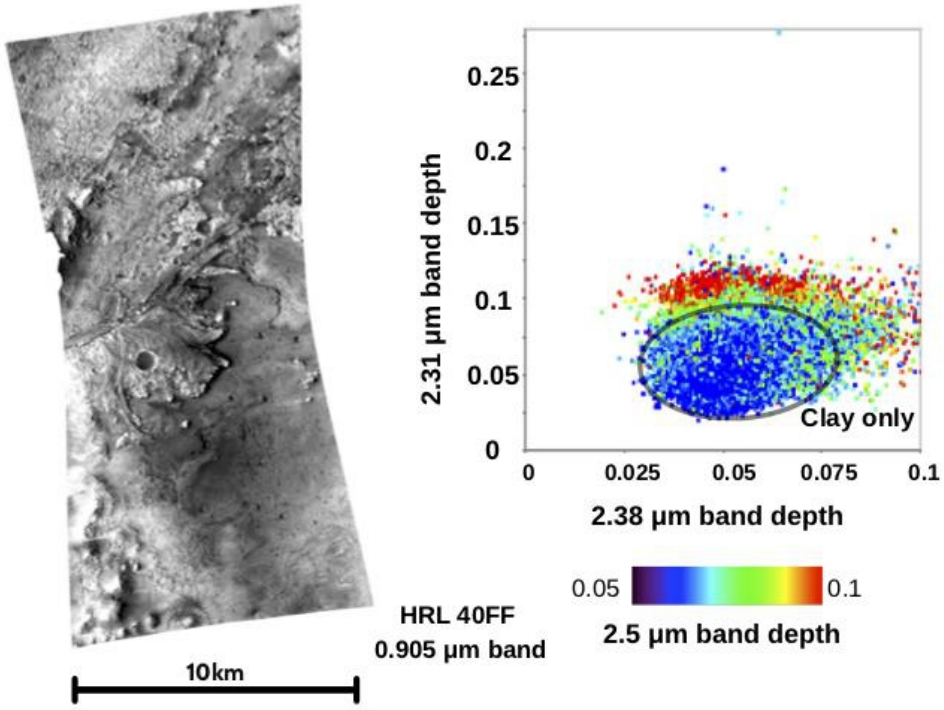


Figure 7. HRL40FF correlation plots of the 2.38 μm band depth (*x*-axis) vs. the 2.31 μm band depth (*y*-axis). Color indicates the 2.5 μm band depth, with red>green>blue.

These correlation plots show when all three bands are present and to what degree. Figure 7 shows that the points that are high in 2.5 μm amplitude are also high in 2.31 μm band amplitude. This allows us to be very specific and make a finding: there are regions of the HRL40FF image that display the presence of all three bands (known as the marginal carbonates, marked on Figure 6). However, there are regions where only phyllosilicates are present, with only a weak to vanishing 2.5 μm carbonate band, indicated by blue colored points. This is marked "Phyllosilicate only" on the Figure 7 correlation plot. This plot makes clear that the highest 2.5 μm band amplitudes are correlated with the highest 2.31 μm band amplitudes.

3.2.3 Venn diagrammatic approach

To get a quantitative estimate of the phyllosilicate and carbonate present with the olivine in this image, we present a Venn diagram populated with data from HRL40FF where olivine, phyllosilicate and carbonate are determined to be present. In order to construct this diagram, we used the following detection rules for the olivine, carbonate, and phyllosilicate bands, which were arrived at iteratively to produce the least noisy CRISM maps, based on previous work (Brown et al. (2020)):

amplitude(olivine 1.0 μm band) > 0.29
amplitude(carbonate 2.5 μm band) > 0.08
amplitude(phyllosilicate 2.38 μm band) > 0.065

For simplicity and reproducibility, these rules were run on the entire georeferenced HRL40FF CRISM image. No attempt was made (at first) to mask the image to regions of interest. We report the border pixels with no data as "edge pixels" and pixels with data as "ground pixels".

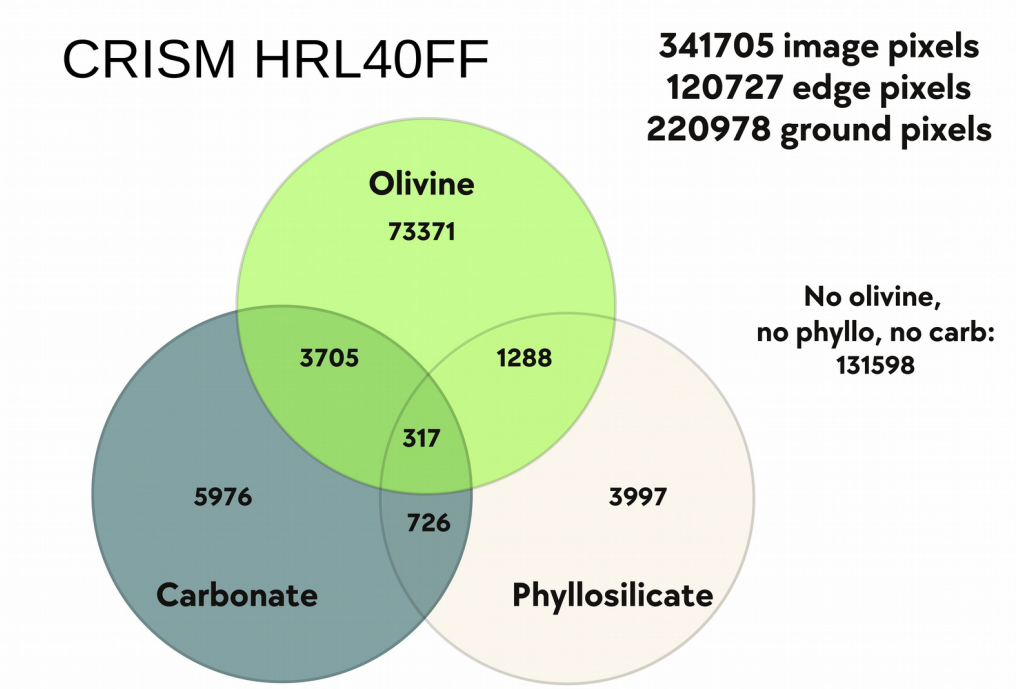


Figure 8. Venn diagram of olivine-phyllosilicate-carbonate detections for HRL40FF showing overlapping pixels containing olivine, phyllosilicate and carbonate. See text for details of calculation.

Figure 8 demonstrates that just more than half of the HRL40FF image is not

olivine, phyllosilicate or carbonate. Roughly one third of the ground pixels have olivine alone detected. “Pure” (i.e. without the other two minerals) phyllosilicate and carbonate are present in roughly 3% and 2% of the pixels, with carbonate being more abundant than phyllosilicate, using orbital data. This plot also demonstrates that there are large amounts of olivine without alteration, and small but significant amounts of phyllosilicate and carbonate without significant olivine signatures in the HRL40FF scene.

In order to take a closer look at the Séítah region as seen from orbit, we have subset the HRL40FF image to a smaller image of only 1750 pixels covering the Séítah region. Figure 9 displays the Venn diagram analysis for just this Séítah subset. This shows that the subset CRISM image mostly contains olivine (~95%), 2% of the pixels contain phyllosilicate and 1.5% are carbonates. Relative to the whole 40FF image, the Séítah subset contains far more pure olivine, but relatively smaller amounts of phyllosilicate and carbonate. This reflects the fact that the Séítah region carbonate and phyllosilicates are relatively difficult to detect from orbit. In the section that follows, we supplement these orbital data with in situ data from the Mars 2020 rover.

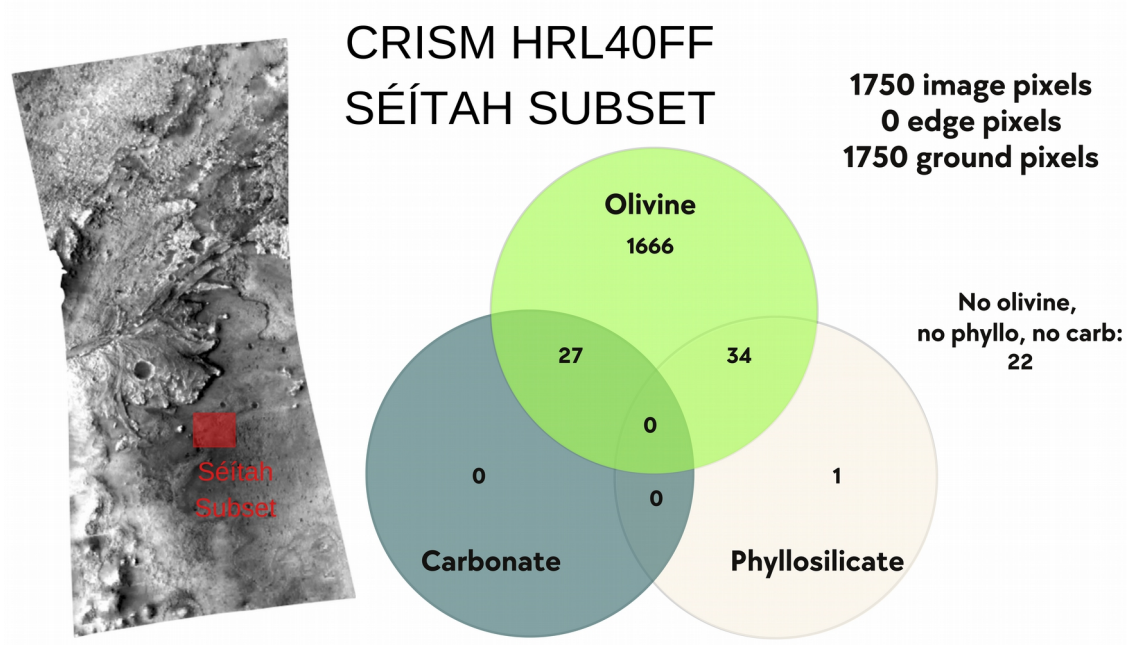


Figure 9. Venn diagram of olivine-phyllosilicate-carbonate detections for the Séítah subset of HRL40FF, which is shown in a red box in the image at left. The diagram on right shows overlapping pixels containing olivine, phyllosilicate and carbonate.

3.3.1 VISIR In situ results

Figure 10 shows the VISIR spectra from the abraded patch Garde taken on Sol 207 from the Séítah formation (Figure 1b). Absorption bands at 2.28, 2.33, 2.38 and possibly 2.46 µm were detected. The lack of a moderate strength 2.5 µm band suggests that this sample is not carbonate bearing; however, see the analysis by Clave et al. (this issue) in which LIBS and Raman techniques suggest small amounts of carbonate are present, and Corpolongo et al. (this issue) in which SHERLOC Raman analyses of the abrasion patch indicate carbonate. The presence of bands at 1.9, 2.28, 2.31, 2.38 and possibly at 2.46 µm in the VISIR indicates that the Garde sample is phyllosilicate bearing. In addition, although weak, there is a band present in the spectra at 2.1 µm, and though this is a little offset (2.09 µm), it is also present in talc. Finally, the 2.46 µm band in serpentine is far weaker than in talc. Figure 10 also shows the example of a spectrum from Dourbes_Tailings_255 which shows the most convincing example we have found of bands at 1.94, 2.28, 2.31, 2.38 and at 2.46 µm, thereby constituting the strongest evidence so far in the Séítah formation for the presence of talc.

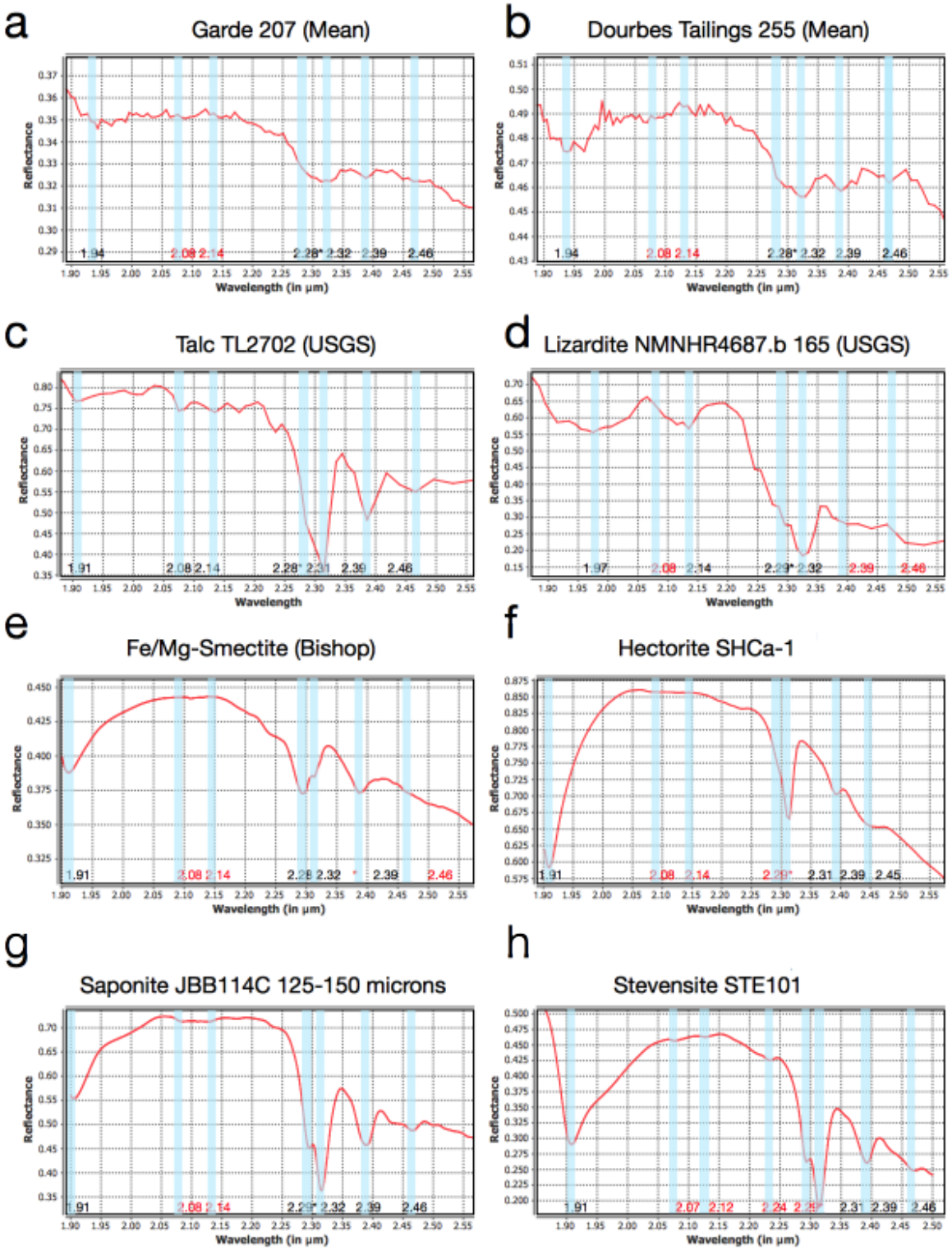
Also shown in Figure 10 are talc, serpentine (lizardite), hectorite, saponite, Fe/Mg smectite and stevensite spectra from Figure 4 for direct comparison with the band positions of the VISIR data.

Table 2 is a summary of the band positions that are present in each of the minerals and targets that are present in Figure 10. Red text indicates a band is weak or missing in this location. An asterisk indicates the presence of a shoulder band.

Table 2. Assessment of SWIR band positions for elements in Figure 10. All values are in µm. Asterisk indicates bands, red text indicates missing or weak band.

Target	1.9-1.97	2.0-2.1	2.1-2.16	2.25-2.29	2.29-2.35	2.35-2.40	2.40-2.5
Garde 207	1.94	2.08	2.14	2.28*	2.32	2.39	2.46
Dourbes Tailings 255	1.94	2.08	2.14	2.28*	2.32	2.39	2.46
Talc TL2702	1.91	2.08	2.14	2.28*	2.31	2.39	2.46
Lizardite NMNHR4867	1.97	2.08	2.14	2.29*	2.32	2.39	2.46
Fe/Mg Smectite	1.91	2.08	2.14	2.28	2.32*	2.39	2.46
Hectorite	1.91	2.08	2.14	2.29*	2.31	2.39	2.45
Saponite	1.91	2.08	2.14	2.29*	2.31	2.39	2.46
Stevensite	1.91	2.07	2.12	2.29*	2.31	2.39	2.47

Based on Table 2, the lack of a 2.46 μm band is somewhat problematic for lizardite and Fe/Mg smectite, however the 2.46 μm band in Dourbes is so weak that it can hardly be disqualifying. At this stage, we see advantages to keeping all 6 candidate minerals in the mix, in the anticipation of spectra with stronger absorbtion bands in the Marginal Carbonates region.


a
Garde 207 (Mean)
b
Dourbes Tailings 255 (Mean)
c
Talc TL2702 (USGS)
d
Lizardite NMNHR4687.b 165 (USGS)
e
Fe/Mg-Smectite (Bishop)
f
Hectorite SHCa-1
g
Saponite JBB114C 125-150 microns
h
Stevensite STE101
Reflectance
Wavelength (in µm)
Wavelength

3.3.2 VISIR, RMI and Mastcam-Z tailings observations

Figure 11 shows the MastCam-Z image for the Dourbes target (abraded rock, Figure 1c), showing the tailings that were captured by the VISIR instrument. It should be noted that the tailings are relatively bright, especially compared with the surrounding bedrock, which is again consistent with the presence of talc or hectorite, which both have high reflectance (>0.9 in the visible region). It should be noted that a decrease of grain size will also increase the albedo, so we do not consider the brightness change to be definitive evidence for talc/hectorite over the other phyllosilicates.

Also shown in Figure 11 are the spectra from the VISIR observations of targets Dourbes, Salette and Brac, which are shown with the spectra of the original rock surface and the tailings after abrasion. The key thing to note, particularly in the Dourbes example, is the relative overall increase in albedo, and the appearance of a band at 2.46 μm which is not present (or only weak) in the original rock. This is strongly suggestive that the tailings have made the presence of the phyllosilicate more obvious in the spectra, and because the 2.46 μm band is not present in lizardite (Figure 10), this also suggests the tailings contain the mineral talc.

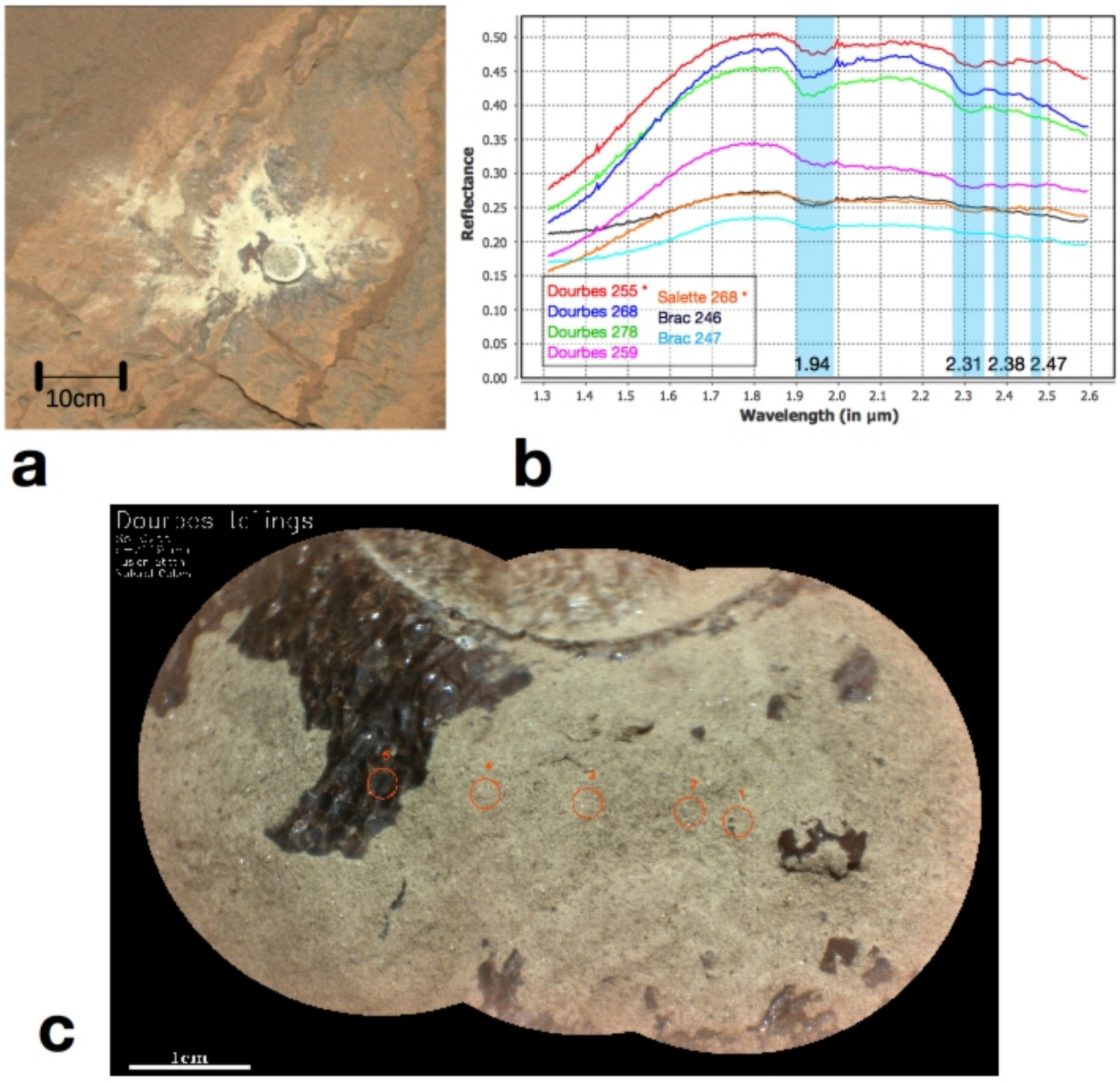


Figure 11. a.) Mastcam-Z of Target Dourbes with tailings on Sol 255 after abrasion in the Brac workspace. Note the brightness of tailings compared to the original rock surface. For scale, the abrasion patch is 5cm across. b.) Spectra of the tailings compared to the original rock surface. Spectra of tailings are marked with an asterisk (*). Note the presence of the 2.46 µm band in the tailings spectra c.) SuperCam RMI image of target Dourbes tailings after abrasion. The numbered red circles indicate the locations at which the data was taken for the Dourbes 255 tailing target and their fields of view (65% of total signal encircled). The blue spectrum in (b) corresponds to the average of the red circles in (c).

3.3.3 SuperCam LIBS ternary plot

Figure 12 shows a Si+Al-Fe+Mg-Ca+Na+K ternary diagram displaying the SuperCam LIBS elemental composition results up to Sol 466 of the mission in black/grey and 10 orange points measured on the target Ubraye_255 which is a rock showing cumulate texture located near the Dourbes abrasion patch in the Brac workspace (Figure 1c). The elemental composition lies very close to the olivine calibration target and also overlaps with the region of talc and serpentine. Given how close these points are and the errors inherent in the planetary LIBS observations (Anderson et al., 2022) it can become challenging to tell these mineral compositions apart by just looking at the major elements. We provide in the supplementary section S3 a discussion of errors and discrimination of points in the ternary plot.

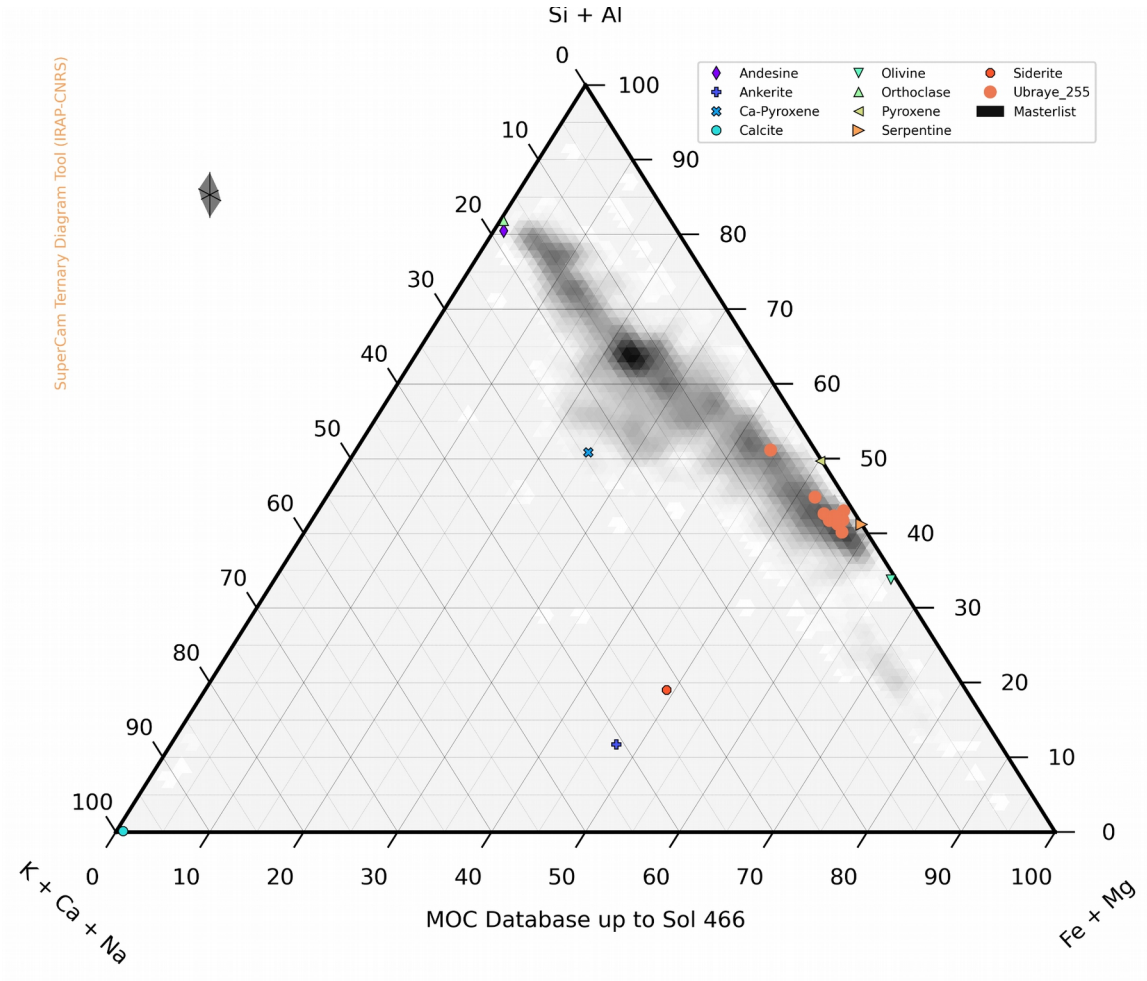


Figure 12. LIBS and geochemistry ternary molar plot indicating Ubraye_255 (orange circles without outline) cumulate target showing overlapping olivine (green inverted triangle) and serpentine (right-pointing orange triangle) standards on board the rover. The gray polygon at top left above the main figure indicates the precision error bars of the plot.

Table 3 gives an example composition from Cine and Ubraye showing the very low (<4 wt %) Al present in these samples. Anderson et al. (2022) Table 5 gives the mean accuracy of the Al_2O_3 estimate at +/- 1.8wt %. For these LIBS targets (Cine/Garde, Udraye/Dourbes which are all in the Séítah region) we can thus eliminate the possibility of Al-bearing phyllosilicates.

Table 3. SuperCam LIBS Elemental abundances from targets Cine_206 (cumulate rock next to Garde in Bastide workspace) and Ubraye_255 (cumulate rock near Dourbes in Brac workspace). Cine_206 point 3 and 5 and Ubraye_255 points 1-4 and 6-10 show a very low amount of Al to within error (<4%) in the cumulate textured rocks. Green boxes highlight the high values of FeO_T and MgO and red boxes highlight the low values of Al_2O_3. S.D. is the standard deviation for that elemental data. Specifically, it is the S.D. of the 25 single shot spectra averaged in each case. All elemental abundance data are in wt %.

Target	#	SiO_2	SiO_2 S.D.	TiO_2	Std Dev	Al_2O_3	Al_2O_3 S.D.	FeO_T	FeO_T S.D.	MgO	MgO S.D.	CaO	CaO S.D.	Na_2O	Na_2O S.D.	K_2O	K_2O S.D.	Total
Cine	1	44.99	1.19	0.43	0.02	3.27	0.53	13.37	1.52	11.54	1.16	15.84	0.68	0.39	0.13	0	0.17	89.83
Cine	2	51.05	1.9	0.04	0	2.94	0.97	25.74	2.06	13.4	1.96	4	0.5	1.19	0.38	0	0.15	98.36
Cine	3	45.13	2.03	0.01	0.02	0.87	0.95	34.59	2.13	28.93	1.27	1.81	1.09	0.37	0.33	0	0.05	111.71
Cine	4	43.39	1.15	0.45	0.03	3.91	1.34	10.43	2.41	11.13	0.92	13.72	0.79	0.49	0.17	0	0.19	83.52
Cine	5	45.34	2.14	0.01	0.01	2.89	1.08	36.37	3.56	28.35	2.22	1.28	0.48	0.51	0.11	0	0.03	114.75

Target	#	SiO_2	SiO_2 S.D.	TiO_2	Std Dev	Al_2O_3	Al_2O_3 S.D.	FeO_T	FeO_T S.D.	MgO	MgO S.D.	CaO	CaO S.D.	Na_2O	Na_2O S.D.	K_2O	K_2O S.D.	Total

Ubraye_255	1	48.72	2.07	0.01	0	2.02	0.88	24.5	2.92	31.5	0.85	1.43	0.54	1.22	0.21	0	0.23	109.4
Ubraye_255	2	44.76	2.2	0.01	0	2.25	0.88	25.94	2.24	27.91	2.7	1.25	0.37	0.34	0.09	0	0	102.46
Ubraye_255	3	51.41	2.57	0.02	0.01	1.57	1.06	23.63	3.05	32.07	0.85	0.8	0.4	1.37	0.17	0.49	0.41	111.36
Ubraye_255	4	45.97	1.52	0.01	0	1.79	0.75	26.35	2.87	28.87	2.03	1.32	0.36	0.3	0.1	0	0.03	104.61
Ubraye_255	5	49.34	2.74	0.04	0	3.92	1.14	28.53	3.78	15.25	2.62	1.65	0.44	1.32	0.14	0.52	0.38	100.57
Ubraye_255	6	46.36	1.4	0.01	0	1.78	1.11	21.35	1.86	30.32	1.16	0.56	0.21	0.29	0.05	0	0.15	100.67
Ubraye_255	7	43.03	7.28	0.03	0.01	2.59	1.32	27.74	3.44	26.59	0.89	1.52	0.42	0.65	0.13	0	0.12	102.15
Ubraye_255	8	47.04	1.74	0.01	0	1.67	0.72	23.18	2.38	29.89	1.04	1.41	0.34	0.65	0.17	0	0.07	103.85
Ubraye_255	9	45.92	2.24	0.49	0.11	3.47	0.79	25.17	2.79	24.79	0.66	1.53	0.3	0.97	0.18	0	0.04	102.34
Ubraye_255	10	45.84	2.18	0.01	0.01	0.54	0.67	27.48	2.68	28.97	2.2	1.46	0.21	0.76	0.15	0	0.03	105.06

4. Discussion

With the insights now gained from bringing together orbital and in situ observations, we now discuss two aspects of the geological history of the olivine-phyllosilicate-carbonate lithology.

4.1 Emplacement of olivine-phyllosilicate-carbonate lithology and exposure today

Flood lavas (e.g. the Columbia River basalt) are characterized by 1) large spatial extent, 2) low viscosity, 3) relatively thick (5-45 m) flows 4) fed by fissure eruptions which are covered by the erupting lava (Plescia, 1990). This is in contrast to (or could evolve over time to) plains lavas (e.g. Indian River basalt), which have smaller-extent, higher-viscosity, thinner flows accompanied by low shields (Plescia, 1990).

Given the low viscosity for the lithology calculated in this paper, the results of RIMFAX (the Ground Penetraing Radar instrument on the Mars 2020 rover) showing the relatively thick flows (Hamran et al. 2022) , and the lack of low shields in the region from orbital mapping, we consider it likely the olivine-phyllosilicate-carbonate at the Séítah formation was emplaced as a flood lava. Future observations at the Marginal Carbonates and outside Jezero will test what other formation mechanisms are involved in other parts of the lithology.

Under the flood lava hypothesis, the olivine-phyllosilicate-carbonate lithology was likely variable in thickness, and in catchment regions of deep relief, such as Jezero crater, the flow formed a lava lake several decimeters thick.

Effect of viscosity. Worster et al. (1993) presented a model of heat flow in a lava lake cooled from above and insulated (or cooled) from below, and utilizing the balancing of the heat equation with crystallization of the melt, they could estimate the total crystallization amount. Their results are shown with their model in Figure 3. Figure 3b also shows that for inferred viscosities typical of the Séítah elemental composition, using the two base cooling scenarios as error bounds, the formation of crystals could have filled up to ~40%+/-10% of the volume of the lake.

Very fluid lava flow episodes have been described in the case of rather recent volcanic events at Central Elysium Planitia (Vaucher et al., 2009) and Tharsis region (Mangold et al., 2010). The combined mineralogical and petrophysical information gained at Jezero with this current study of Séítah unit points to the fact that low viscosity lava flows did occur, possibly at large scale, early in the martian history; as such, it may place valuable constraints on the ultramafic composition of the Martian interior at a rather early to intermediate stage of chemical differentiation.

Depth of lava lake. We suggest that the lava lake has been eroded back to reveal the cumulate portion of the lava lake and then covered by crater fill consisting of igneous material from later eruptions such as the Máaz formation (Udry, this issue) and Artuby member of the Máaz unit (Alwmark, this issue) which has itself then been eroded back in locations to reveal the olivine-phyllosilicate-carbonate beneath it (Hiesinger and Head, 2004; Goudge et al., 2015). Ground-penetrating radar imaging performed by RIMFAX in the vicinity of Séítah reveal sub-horizontal high-amplitude reflectors down to 15 m burial depth including lenticular reflector geometries interpreted to most likely represent cross-sections of igneous flow lobes (Hamran et al., 2022) The base of the Séítah unit currently remains unclear, but is at least ~15 m below the surface at the deepest locations observed so far by RIMFAX.

Assuming for the sake of the argument that the Séítah formation was ~10 m deep if it was formed as a lava lake, this suggests (using Figure 3b) that approximately the lower ~4+/-1m consists of the cumulate Bastide workspace (Figure 1b) in which Cine_206 and Garde_207 were observed.

4.1.1 Spatial distribution of lithology as flood lava

Wilson and Head (1994) pointed out that lava flows on Mars will extend roughly six times further than lava of the same composition on Earth. as mentioned in Section 2, this is due to the lower gravity and the lower pressure of the current Martian atmosphere. A higher pressure Noachian atmosphere would decrease the runout. The emplacement of the lava that created the Séítah formation olivine cumulate covered a significant portion of the planet (Figure 13). Kremer et al. (2019) mapped the unit to extend over both sides of the Nili Fossae, and showed the unit to be predominantly exposed on the eastern side of the Nili Fossae, and truncated by Syrtis Major to the south, and later cratering activity, most notably by Hargraves Crater in the middle of the two predominant Fossae. Considering only the region of the Isidis Planitia basin complex currently covered by the olivine-phyllosilicate-carbonate unit, and excluding the olivine on the Libya Montes region on the southern edge of the basin, this area extends roughly 500 km north to south and 350 km east to west and is approximately the size of Florida (in 2022) in the United States (~170,000 km^2).

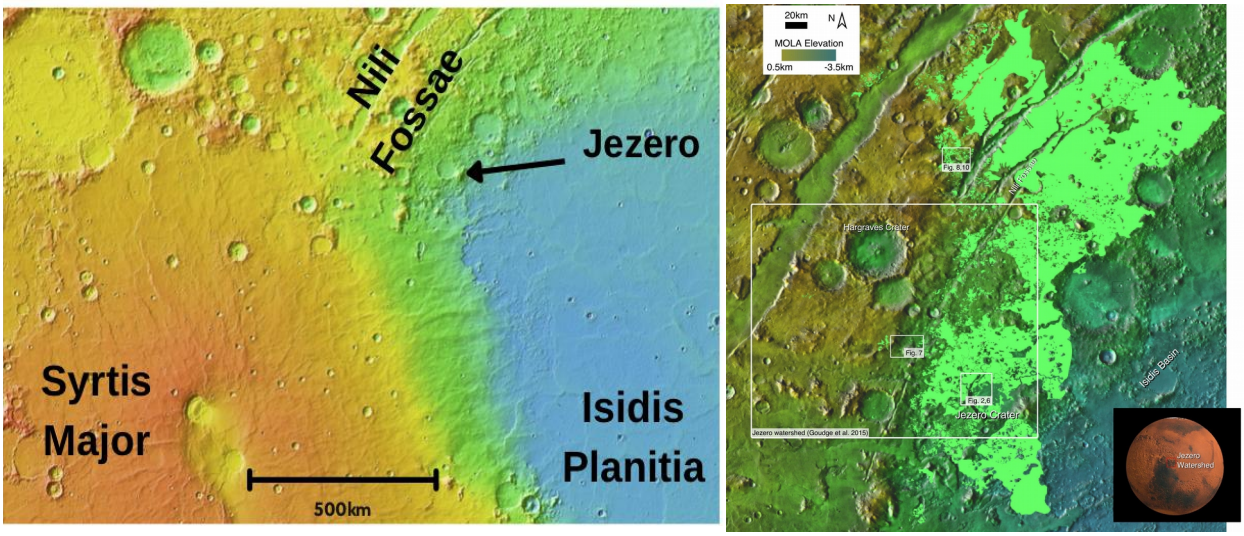


Figure 13. a.) MOLA image showing the relationship of western edge of the Isidis Basin complex, giving the relative positions of Syrtis Major, Nili Fossae and Jezero Crater. b.) Figure 1 of Brown et al. (2020) showing the extent of the olivine-phyllosilicate-carbonate lithology as mapped by Kremer et al. (2019) in a region about the size of Florida. Note obscuration by later crater ejecta blankets, especially Hargraves Crater.

The direct contact between the Syrtis Major lavas and the olivine-phyllosilicate-carbonate is striking and leads one to conjecture whether the Fossae extends beneath the Syrtis Major construct; although this has not been previously addressed, we will now postulate about how this geophysical relation might have occurred. Rampey and Harvey (2012) identified the wide depressed footprint of

Syrtis Major and suggested that:

> "early eruptions at Syrtis Major might have consisted of voluminous low-viscosity, possibly komatiitic magmas, whose powers of thermal erosion … may have been responsible for the wide, depressed footprint of Syrtis Major."

And at the end of the same paragraph:

> "... the youngest and most visible surface rocks of which would be basaltic komatiites, but whose earlier, more voluminous rocks would have been ultramafic. In the absence of tectonic effects these deep-seated rocks would have restricted exposure."

We would contend that the missing earlier, more voluminous ultramafic rocks, are, in fact, the olivine-phyllosilicate-carbonate lithology we have been studying. Somewhat more speculatively, we might make some passing comments which may be useful for the collective consciousness to test in future investigations. The suggestions are four-fold and currently have no basis in observation, but are consistent with observations so far and may be tested in future observations, for example when the *Perseverance* rover can inspect the Nili Fossae in situ.

1) The fissures that were the source of the olivine-phyllosilicate-carbonate lava were possibly linked to, or in fact used, the Nili Fossae fractures on the northwest of Isidis Planitia, and opened after the Isidis impact (Wichman and Schultz, 1989). The analogy might be made to the relationship between Cerberus Rupes and Cerberus flood lavas (Plescia, 1990) in Central Elysium Planitia.

2) At Cerberus, it is possible that the flood lava evolved to a plains style lava. This may also have happened in the Isidis basin, where Syrtis Major may have been created after fissure lavas became more viscous, and the Syrtis shield may now cover fissures from an earlier phase of flood volcanism that produced the olivine-phyllosilicate-carbonate lithology. Figure 13 shows a Mars Orbiter Laser Altimeter (MOLA) altimetry map of the region highlighting the spatial relationships between the Isidis, Syrtis, Nili Fossae and Jezero crater. Rampey and Harvey (2012) suggested that Syrtis Major evolved from ultramafic to mafic over time and that super-hot komatiite lava may have eroded basement rocks. As noted by previous authors, no olivine has been found to date in the Fossae floor or walls (Hamilton and Christensen, 2005; Mustard et al., 2007). However, it is possible that evidence of this olivine may yet be found buried at depth by Syrtis Major lavas. Channelized lava flows travel much further than non-channelized flows (Pinkerton and Wilson, 1994), and the last Syrtis Major lavas may have traveled more easily in the Nili Fossae graben and subchannels.

3) At Cerberus, the lavas are much younger than the olivine-phyllosilicate-carbonate, post-date water-related fluid channels, and are often guided by those

channels. Volcanic activity commenced at Syrtis between 3.6-3.9 Ga (Robbins et al., 2011) around the same time as the oldest deposits on Tharsis were emplaced at 3.67 Ga (Isherwood et al., 2013). The age of the olivine-phyllosilicate-carbonate (3.82 +/- 0.07Ga, Mandon et al. (2020)) would imply that no surface water channels were available to confine the flow, but might instead be governed by the widespread emplacement of the Noachian basement rocks.
4) During emplacement of this super-hot lava lake, it might be possible that Séítah sank into and eroded the regolith it was laid down upon (Huppert et al., 1984).
Evidence for all four of these suggestions might be found in exposures sounded by RIMFAX during the Mars 2020 mission in the Outside Jezero part of the mission.

4.1.2 Properties of cumulate layer and terrestrial analog

Huppert and Sparks (1981) discuss the manner in which an olivine bearing ultramafic cumulate layer might form in a magma body beneath a less dense plagioclase bearing basaltic layer. This model is a reasonable candidate for the formation of the Séítah formation and the Artuby member of the Máaz formation (see also Crumpler et al., this issue). It may even be possible that the lava which formed Séítah was contaminated during eruption and emplacement (Huppert and Sparks, 1985), and given its low viscosity and high temperature, this may have differentiated the upper Issole member as a contaminated version of the lava producing the lower cumulate Bastide member (Crumpler et al., this issue).

Figure 14 compares the Martian olivine cumulate target Cine_206 and the terrestrial olivine cumulate target AJB0503100 from the Dresser Formation in the Pilbara Craton of Western Australia. Both show equant (1 to 1.5 mm diameter) grains that are interlocking. Both were likely formed as the basal part of a lava flow or ponded lake when the olivine crystals fell out of suspension and collected at the basal cumulate layer. The terrestrial example on the right was sampled from the talc-carbonate altered unit at the bottom of the Mt Ada Basalt in the Dresser Formation of the North Pole Dome in the Pilbara region of Western Australia. The sample is part of a komatiite sequence and specifically the B2 adcumulate layer (Arndt et al., 2004). We have shown in this paper that the target Cine shares many spectral, mineralogical and morphological characteristics of the terrestrial target.

Although a Martian lava lake is likely to have formed from a cooler mantle source than the terrestrial komatiite (Putirka, 2016), we contend that it is the viscosity we have derived here that governs emplacement properties of the lava flow. Direct comparison of the viscosity of the two units is challenging due to the high Mg nature and alteration of the terrestrial sample (MgO 18.9 wt.%, FeO_T 7.6 wt.% from microprobe, presented in Brown (2006)) versus the higher Fe content of Target Cine (SiO_2 45.24 wt %, MgO 28.64 wt.% and FeO_T 35.48 wt.%, which is the average of Cine points 3 and 5 from Table 3). It should be noted that both these rocks do not have a true bulk rock content because they are cumulates of material that were sourced from elsewhere, most likely within the upper zone of the lava lake.

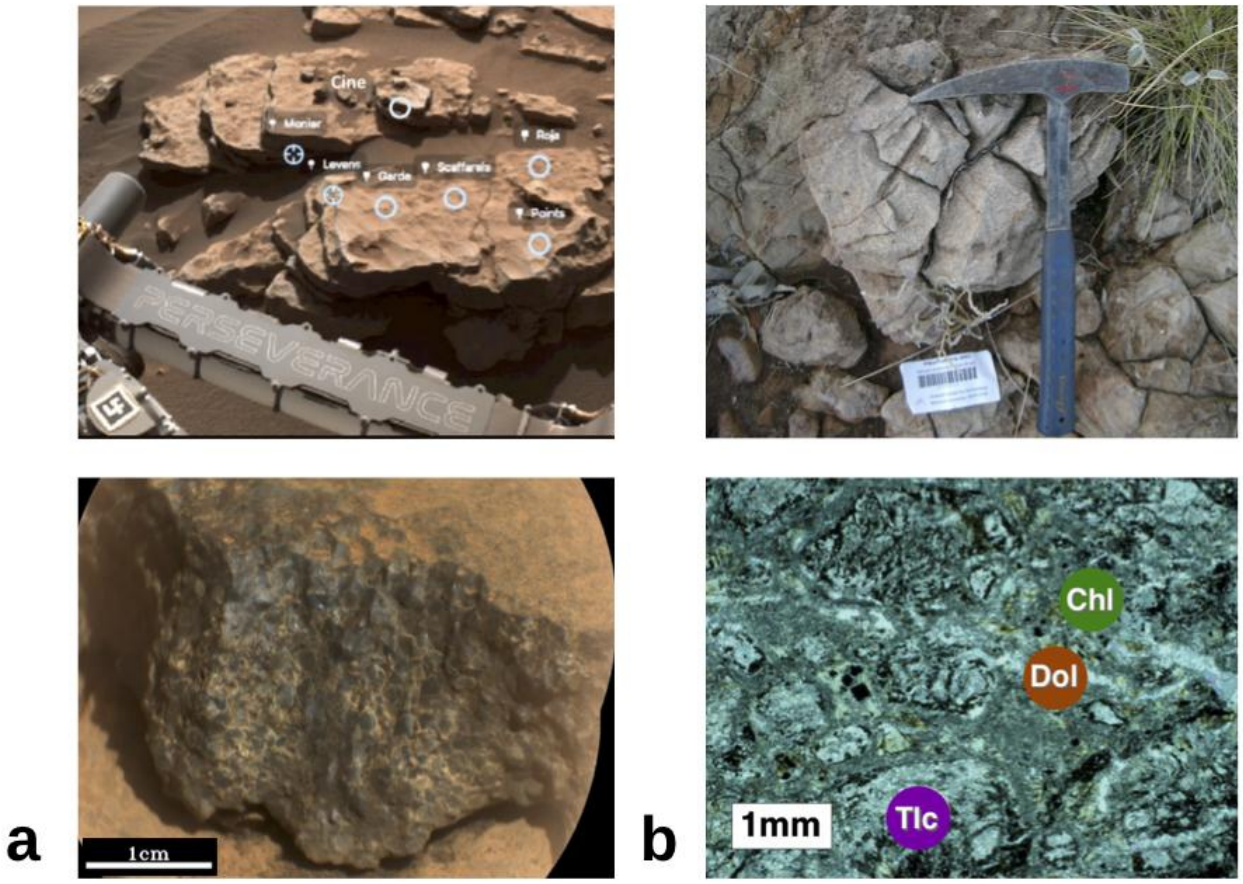


Figure 14. a.) (bottom) SuperCam RMI of target Cine showing olivine cumulate texture at mm scale. (top) Bastide workspace, location of Cine relative to Garde, rover arm for scale. b.) (bottom) Thin section of talc-carbonate sample AJB0503100 from Brown (2005; 2006) with mm size talc (Tlc) replacing olivine, dolomite (Dol) and chlorite (Chl) identified using electron microprobe. (top) Context image for target AJB0503100 with hammer for scale.

4.2 Post emplacement phyllosilicate-carbonate alteration

The fact that ~3.94 Ga olivine is still well preserved on Mars and relatively unaltered speaks volumes to the amount of alteration and volatiles present on the fourth planet. In terrestrial greenstone Archean terranes, olivine is usually only present as a polymorph of its former self, often replaced by phyllosilicates such as serpentine or talc (Figure 14b).

Wilson and Head (1994) suggest that lavas on Mars have fewer volatiles than lavas on Earth. Despite this, the low amount of alteration present means it is still possible that the alteration fluids were juvenile water and CO_2, sourced from the fissure, as suggested and discussed by Brown et al. (2020).

The limited extent of post emplacement phyllosilicate and carbonate alteration is evident from the Venn diagrams we have presented in Figures 8 and 9, and is also the subject of two studies, including phyllosilicate (Mandon et al., this issue) and carbonate (Clave et al., this issue). Mandon et al. (2022) show that in olivine-carbonate-serpentine mixtures, the bands of serpentine are deep for percentages of < 5 wt.% of serpentine. This also has implications for the alteration history, as the rocks in the Séítah formation might not have extensively interacted with fluid compared to some other parts of the regional olivine unit, which appear more altered from orbit.

4.3 Other emplacement models

There are a number of aspects of the regional morphology and properties of the olivine-phyllosilicate-carbonate lithology that remain poorly understood, including draping morphology and laterally continuous layering and lack of clear lobate features. These have been used to argue in favor of an explosive volcanic origin for the lithology (Rogers et al., 2018; Kremer et al., 2019; Mandon et al., 2020; Ruff et al., 2022). Ravanis et al. (2022) report on modeling to determine grain sizes for Martian Nili Fossae ashfall deposits in order to better understand the potential for pyroclastic fall deposits to be emplaced away from the eruption source. Wilson and Head (1994) modeled clasts with diameters between 100 μm and 10 mm and maximum travel ranges of pyroclasts in current day (6 mbar) Martian atmosphere for eruption cloud heights from 50-300 km. Their Figure 21 demonstrated that the maximum range drops sharply as grain size increases. They also presented a modeled maximum pyroclastic flow run out length as a function of eruption velocity for three different surface friction coefficients. Their Figure 23 shows that the 1.5 mm olivine grain size would only travel about 20 km away from the vent through pyroclastic fall, and thus far, the Mars 2020 team is yet to discover

pyroclastic signatures (e.g. fiamme) in the imagery of abrasion patches obtained by the rover.

4.4 Future prospects and Mars Sample Return

At the time of writing this paper and other accompanying papers in the Jezero Crater Floor collection, the ongoing progress of the *Perseverance* rover has brought it to the Jezero Delta campaign of its mission. Within the delta, the delta curvilinear unit (*DCu*) and some parts of the delta have been documented to contain carbonates, from orbital studies (Goudge et al., 2015, 2018), and are now being characterized by surface observations (Mangold et al., 2021). We anticipate unexpected findings in the rocks and cobbles of the delta, which will provide us views of the outside watershed at irregular intervals. Figure 15 gives the timeline of events in Martian geological history that are critical to understanding its evolution and reliance upon regional and global events. We point out that this timeline shows an approximate correlation between the Jezero Delta and the recently hypothesized second generation ocean, called Deuteronilus, aged approximately at 3.65Ga, and discussed in Citron et al. (2018) which they hypothesized filled Isidis with water. The existence and timing of these oceans is a source of ongoing debate (Sholes et al., 2021; Sholes and Rivera-Hernández, 2022) and during the Delta campaign we may hope to deepen our understanding of the source of the water that once filled Jezero.

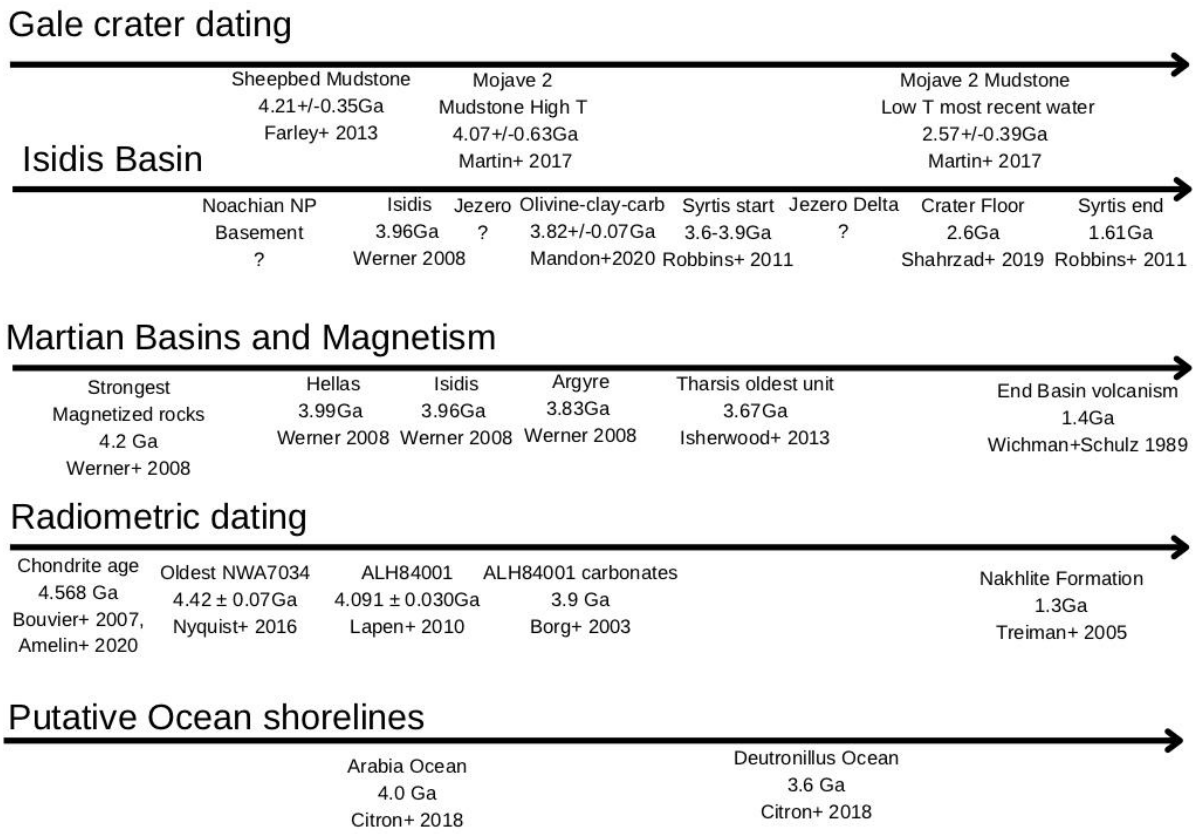


Figure 15. Timeline of events relevant to the emplacement and retention of the olivine-phyllosilicate-carbonate lithology. Dates are from various sources (Amelin et al., 2002; Borg et al., 1999; Citron et al., 2018; Farley et al., 2013; Isherwood et al., 2013; Lapen et al., 2010; Mandon et al., 2020; Martin et al., 2017; Nyquist et al., 2016; Robbins et al., 2011; Shahrzad et al., 2019; Treiman, 2005; Werner, 2008; Wichman and Schultz, 1989). NP = Nili Planum

After the Delta campaign, the rover will pass either nearby or through the marginal carbonates, which are the greatest concentration of carbonates at Jezero (Figure 6).

Throughout its mission *Perseverance* is collecting samples for later return to Earth as part of the Mars Sample Return Mission. The samples collected in the Crater Floor campaign are discussed in Simon et al. (this issue). Upon return to Earth, we anticipate these activities to test the postulates and expand the findings of this paper:

* Use of radiometric age dating on the grains and mesostasis,
* Determination of phyllosilicate and carbonate type using Microprobe and X-ray Diffraction,
* Determination of phyllosilicate and carbonate relative timing using X-ray tomography,
* Source of carbonate using stable isotopes,

* Age dating of delta topsets and bottom sets samples to constrain the age of the Jezero Delta.

5. Conclusions

The *Perseverance* mission completed the Crater Floor science campaign in April 2022, and thus the observations and interpretations we have reported here are limited to the olivine-phyllosilicate-carbonate as it appears at the Séítah formation. We can infer characteristics of the unit beyond Jezero based on orbital observations (Goudge et al., 2015; Horgan et al., 2020; Mandon et al., 2020; Brown et al., 2020). For example, one observation of the orbital data demonstrated that the largest olivine spectral features (and inferred grain size) are found outside, and to the north, of Jezero (Brown et al., 2020). However, our findings for this regional lithology will remain uncertain until we are able to reach these locations for in situ observations. These occasions will no doubt be as informative as the visit of *Perseverance* to Séítah.

We have made the following new findings and interpretations:

1. We have used a Venn-diagrammatic approach to establish that the previously identified "olivine-carbonate" lithology should actually be called the "olivine-phyllosilicate-carbonate" lithology.
2. Again using the Venn-diagrammatic approach, we have demonstrated the limited extent of alteration by phyllosilicate and carbonate, particularly for the Séítah subset of HRL40FF.
3. We propose that the olivine-phyllosilicate-carbonate lithology at the Séítah formation was emplaced as a flood lava, with low viscosity, relatively deep flows, and lack of low shield volcanoes.
4. We have presented a model suggesting that thermal-convection-driven cooling, balanced by crystallization of the lava, may have taken place in order to form large mm-sized crystals seen in the Séítah formation.
5. We have presented a model where flood lava ponded into a lava lake in Jezero crater, and then 40+/-10% of the magma in the lake crystallized. Based on RIMFAX data suggesting the Séítah formation was approximately 10m thick, this would lead to the bottom ~4m of the Séítah unit consisting of the Cine and Garde Bastide workspace olivine cumulate outcrop sampled by *Perseverance*. The original magma body would necessarily have been thicker to allow differentiation of the olivine crystals and settling to the bottom of the chamber. This upper part would be more pyroxene than

olivine-rich, as described in Wiens et al. (2022). It is likely that this portion of the magma lake was removed prior to emplacement of later lava flows.

6. As seen in Figure 6, there are regions of the CRISM HRL40FF image that display the presence of all three olivine, phyllosilicate, and carbonate bands (known as the marginal carbonates). However, there are regions where only phyllosilicates are present, with only a weak to vanishing 2.5 μm band. This is marked "Phyllosilicate only" on the Figure 7 correlation plot.
7. As seen in Figure 4 and 10, the IR spectra of phyllosilicates talc, lizardite, Mg/Fe smectite, hectorite, saponite and stevensite, are broadly consistent with the spectra of Garde 207 and Dourbes Tailings 255. As seen in Figure 10, we have shown that talc is a strong match for the phyllosilicate signal in the VISIR spectra for Garde, and especially the Dourbes tailings. We regard the suggestion of talc to be only plausible or tentative until the VISIR calibration is further improved (Royer et al., this issue), or more observations are made in locations of higher phyllosilicate abundance, for example in the Marginal Carbonates region or potentially in the Noachian region outside of Jezero.
8. Based on the lack of a 2.28 μm band, we can eliminate nontronite and hisingerite as the phyllosilicates in Gardes and Dourbes.
9. Based on Table 3, showing the LIBS results for Cine/Garde and Ubraye/Dourbes, the low aluminum amounts for the olivine cumulate are not consistent with the presence of Al bearing phyllosilicates.
10. We have formulated a new conjecture that the olivine-phyllosilicate-carbonate lithology is the missing earlier, more voluminous ultramafic rocks proposed by Rampey and Harvey (2012). Needless to say, this will require further testing in situ.

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818602.

Open Research

All CRISM and VISIR data used can be obtained from the Planetary Data System (PDS) Geoscience Node (http://ode.rsl.wustl.edu/mars). All data and software used to produce the datasets in this manuscript are available at the Harvard Dataverse (https://doi.org/10.7910/DVN/QEIVES).

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