# RECENT LIQUID WATER ON MARS INFERRED FROM SHOCK DECOMPOSITION ANALYSIS OF PHYLLOSILICATES WITHIN IMPACT CRATERS

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**Introduction:** Multiple phyllosilicate deposits on Mars appear associated with impact craters and distributed throughout the rims, ejecta and central peaks [1-4]. It has been suggested that these deposits derived from pre-existing phyllosilicate-rich materials that were excavated in the process of crater formation [3,5], and are therefore indicative of aqueous activity early in the history of the planet. We have analyzed if the distribution of phyllosilicates within impact craters is consistent with their stability against shock and thermal decomposition, to test the hypothesis that these deposits represent vestiges of early Mars aqueous activity.

**Thermal and shock stability:** Hydrated silicate minerals are characterized by a low thermal stability due to the presence of adsorbed water and OH-groups in their lattice, and many mixed-layer phyllosilicates become unstable at temperatures of ~370 K [6]. Dehydroxilation starts in montmorillonite at temperatures as low as 500 K [7], nontronite is completely decomposed at 625 K [8], the iron-enriched smectites dehydroxylate at 725 K [9], and kaolinite starts undergoing dehydroxylation at 775 K [10]. In general, 920 K can be considered the upper average decomposition temperature for phyllosilicates, as it is the decomposition temperature of serpentine to glass and anhydrous silicates [11]. The large shock pressures resulting from a meteor impact can also induce the dehydration of silicate minerals. Theoretical estimates and shock recovery experiments show incipient to complete water loss from 200 to 600 kbar respectively, with an average 40% water loss at 300 kbar shock pressures [12]. Up to 2/3 of the interlayer water of nontronite samples is lost at shock pressures of 300 kbar [13], and phyllosilicates shocked above 260 kbar can be changed to an almost amorphous state [14].

We adapted the model derived by Kieffer and Simonds [15] to compute the shock-induced pressure and the residual temperature of an impact event on Mars. The top panel of Figure 1 shows the radii of the areas heated up to 370 K and 920 K and shocked at 300 kbars around the impact point. The radius of the area experiencing temperatures and pressures leading to complete phyllosilicate decomposition, the shock decomposition area (SDA), is ~1/5 of the crater radius at the surface level. Subsurface clay-bearing horizons located in the central peak and surrounding areas, would not survive the impact process. However, peripheral deposits outside these regions could survive the impact forces, due to shock pressure and temperature dissipation.

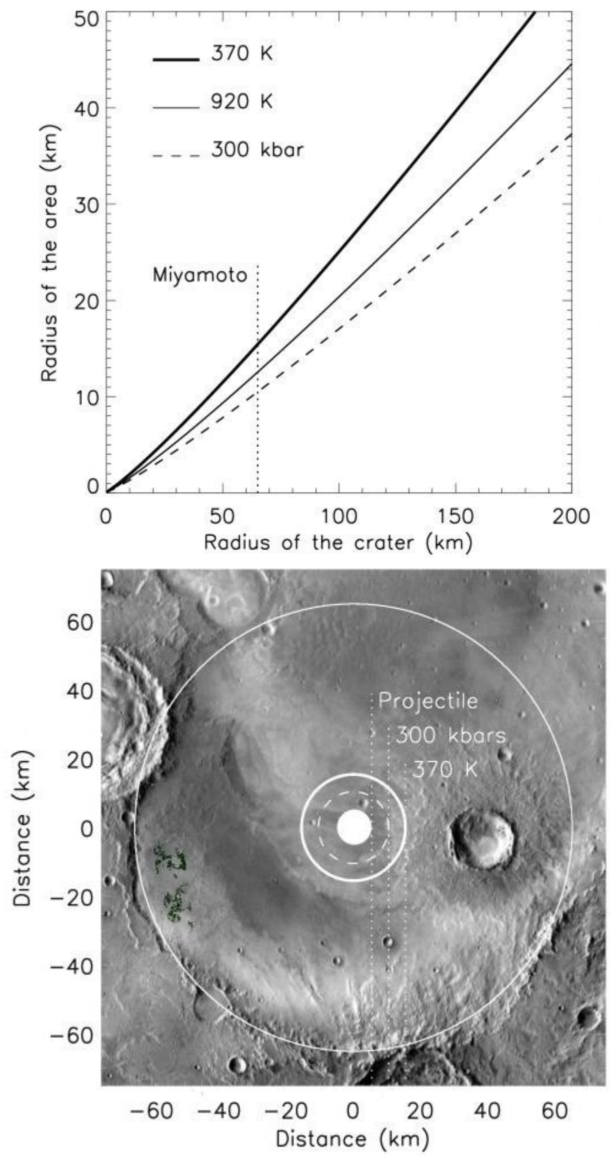


**Figure 1.** Top: Radius of non-stability areas for phyllosilicates, as function of the crater diameter, due to a shock-induced pressure (300 kbars) and two different temperatures (370 and 920 K). Bottom: Simulation for Miyamoto crater. The phyllosilicates (green) lie outside the SDA.

In the bottom panel of Figure 1 we simulate the case of Miyamoto crater in Meridiani Planum. The green deposits, in the southwest portion of the crater, exhibit spectral features indicative of Fe/Mg-smectite clays [16], and are located in the periphery of the crater floor, outside the SDA. Therefore their presence is compatible with excavation models of crustal phyllosilicate-rich deposits formed prior to the impact.

**Late stage phyllosilicates:** We have analyzed several examples of impact craters with phyllosilicate deposits that fall within the SDA. The crater shown in Figure 2 is 40 km in diameter and 2 km depth, and is located close to the northern edge of the Syrtis Major Volcanic Plains (SMVP). It is deep, bowl-shaped with maximum slopes of 25º, and has a well developed central peak with a central pit, fresh ejecta blankets, megabreccia on the crater floor, terraced walls, and a well-defined and complete rim. These features are all diagnostic of a fresh impact [17]. The formation age of the SMVP has been constrained to the Hesperian [18], and this places an upper limit for the age of the impact crater. The crater is located in the Late Hesperian units and there are only meter-scale superposed impacts on the crater floor, and two larger impacts (~2 km diameter) on the ejecta. This suggests that the crater formed towards the end, or after the volcanic activity that resulted in the formation of the SMVP.

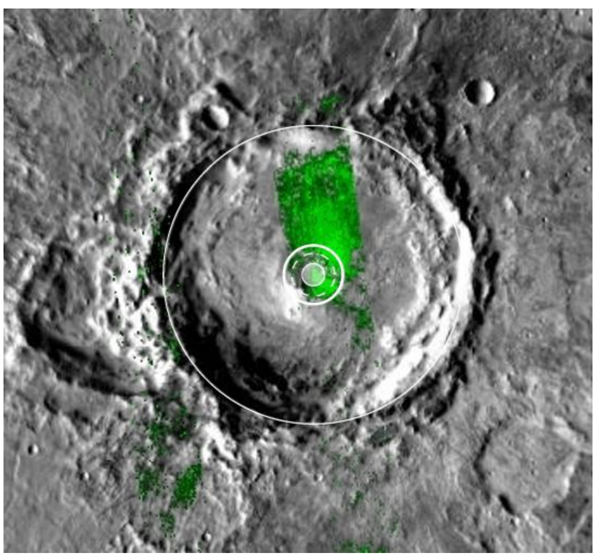

**Figure 2.** Crater in the Syrtis Major region (71.8E, 17.0N). Areas indicated in green have spectral signatures of phyllosilicates. The SDA is delimited by circles in center of crater.

Hydrated silicate deposits, mainly composed of Fe/Mg smectites and chlorites appear inside the SDA. At high resolution, the deposits have the appearance of bright outcrops, partially covered with sand dunes. Current models suggest that these deposits are Noachian in age and were excavated after the impact [3,5]. But phyllosilicate outcrops appear on the rim of the central peak, where our model predicts a SDA, and Mg/Fe smectites would dehydroxylate and decompose completely [9,12,13]. Our model also predicts that any preexisting phyllosilicate deposits would be completely decomposed down to a depth of ~2 km beneath the SDA.

Phyllosilicates in this crater appear to have form through hydrothermal activity associated with the impact itself, as is documented on Earth [19]. Impact-induced hydrothermal activity is initiated by an impact event into water-rich or ice-rich crustal material. The resulting release of heat provides a thermal driver for the circulation of water and volatiles [20, 21]. Numerical models of groundwater flow within freshly formed martian impact craters indicate that hydrothermal systems will develop if sufficient water is present [22]. Impact triggered hydrothermal activity can last over $10^5$ yrs for a crater the size of that analyzed here [21]. Based on this cumulative information, we favor the hypothesis that the presence of phyllosilicates within the SDA of this fresh impact crater is indicative of hydrothermal activity on Mars after the Noachian.

**References:** [1] F. Poulet, et al., Nature 438, 623-627 (2005). [2] J. P. Bibring, et al., Science 312, 400-404 (2006). [3] J. F. Mustard, et al., Nature 454, 305-309 (2008). [4] J. L. Bishop, et al., Science 321, 830-833 (2008). [5] S. M. Pelkey, et al., J. Geophys. Res. 112, E08S14, doi: 10.1029/2006JE002831 (2007). [6] K. Byrappa, M. Yoshimura, Handbook of hydrothermal technology (William Andrew Inc., 2001), 870 pages. [7] R.E. Milliken, J. F. Mustard, J. Geophys. Res., 110, doi:10.1029/2005JE002534 (2005). [8] P. Gavin, et al., LPSC 2007. [9] R. L. Frost, et al., Thermochim. Acta 346, 63-72 (2000). [10] H. He, et al., J. Am. Ceramic Soc. 88, 1017-1019 (2005). [11] G. W. Brindley, J. Lemaitre, in Chemistry of clays and clay minerals (ed. A. C. D. Newman); Mineral. Soc. Monogr. 6, pp. 319-370, Longman (1987). [12] M. A. Lange, et al., Geochim. Cosmochim. Acta 49, 1715-1726 (1985). [13] M. B. Boslough, et al., LPSC 1980. [14] J. Akai, T. Sekine, Proc. NIPR Symp. Antarct. Meteorites 7, 101-109 (1994). [15] S. W. Kieffer, C. H. Simonds, Rev. Geophys. Space Phys. 18, 143-181 (1980). [16] S. M. Wiseman, et al., LPSC 2008. [17] R. E. Arvidson, Icarus 22, 264-271 (1974). [18] H. Hiesinger and J. W. Head III, J. Geophys. Res. 109, doi:10.1029/2003JE002143 (2004). [19] M. V. Naumov, Geofluids 5, 165 – 184 (2005). [20] H. E. Newsom, Icarus 44, 207-216 (1980). [21] O. Abramov, D.A. King, J. Geophys. Res. 110, doi:10.1029/2005JE002453 (2005). [22] J. A. Rathbun, S.W. Squyres, Icarus 157, 362-372 (2002).