



# Post-Noachian Water Activity on Mars Inferred from Shock Decomposition Analysis of Phyllosilicates within Impact Craters

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# The Rationale

- Meteorite impacts can generate pressures in excess of 4000 kbars and temperatures in excess of 1000 K (Kieffer and Simonds, 1980)
- Phyllosilicates dehydroxylate (partially or completely) at temperatures  $> 700$  K, and are destroyed at temperatures  $> 1000$  K (Brindley and Lemaitre, 1987, Gavin et al., 20 minutes ago).
- 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 (Lange et al. 1985)
- A theoretical Total Destruction Zone (TDZ) can be defined for impact craters as a function of T and P. Within the TDZ pre-existing phyllosilicates would not have survived the impact temperatures and pressures.
- Phyllosilicates within the TDZ must post-date the impact event.

# The Model

- Off-the-shelf model based on Kieffer and Simonds (1980) and Shoemaker and Wolfe (1982). Updated by Michael Carr (1989) to study impact-induced release of CO<sub>2</sub> on Mars.
- Provides a VOLUME around the impact point that has been shocked to a Pressure  $P$  and heated to a Temperature  $T$ .

# The Model

- Kinetic energy (W):

$$D = \left( \frac{g_e}{g_m} \right)^{\frac{1}{6}} c_f k_n W^{\frac{1}{3.4}}$$

D is the crater diameter,  $g_e$  and  $g_m$  are the gravitational acceleration of Earth and Mars,  $c_f = 1.3$  is the collapse factor,  $k_n = 0.074 \text{ km kiloton}^{-1}$ , and W is the kinetic energy

- The kinetic energy gives the meteorite radius once its density and impact velocity are established. We assume a meteorite density of  $2.98 \text{ g cm}^{-3}$ , equivalent to basalt, and an impact velocity ( $v_i$ ) of  $15 \text{ km s}^{-1}$

# The Model

- Initial Pressure at impact:

$$P - P_0 = \rho_0 U_t u_t$$

where  $P_0$  is the pressure before impact,  $\rho_0$  the density of the target, and  $U_t$  and  $u_t$  the shock and particle velocities, respectively

- Particle velocity: 
$$u_t = \frac{2c_0(1 + \nu_i)}{\nu_i(c_0 + s\nu_i)}$$
- Shock wave velocity: 
$$U_t = c_0 + su_t$$

# The Model

- The fraction of energy transferred to the target (f) is:

$$f = 1 - \frac{[(v_i - u_t)^2] + u_t^2}{v_i^2}$$

- The energy transferred to the target can be used to determine the radius of the target at pressure P ( $r_0$ ):

$$r_0 = \left( \frac{v_i}{u_t} \right)^{\frac{2}{3}} \left( \frac{3}{2} f \right)^{-\frac{1}{3}}$$

- A peak shock (P) decreases from the point of impact with a -3.6 decay law (Kieffer & Simonds, 1980). The radius of the area ( $r_1$ ) around the point of impact shocked at a pressure greater than  $P_1$  is therefore

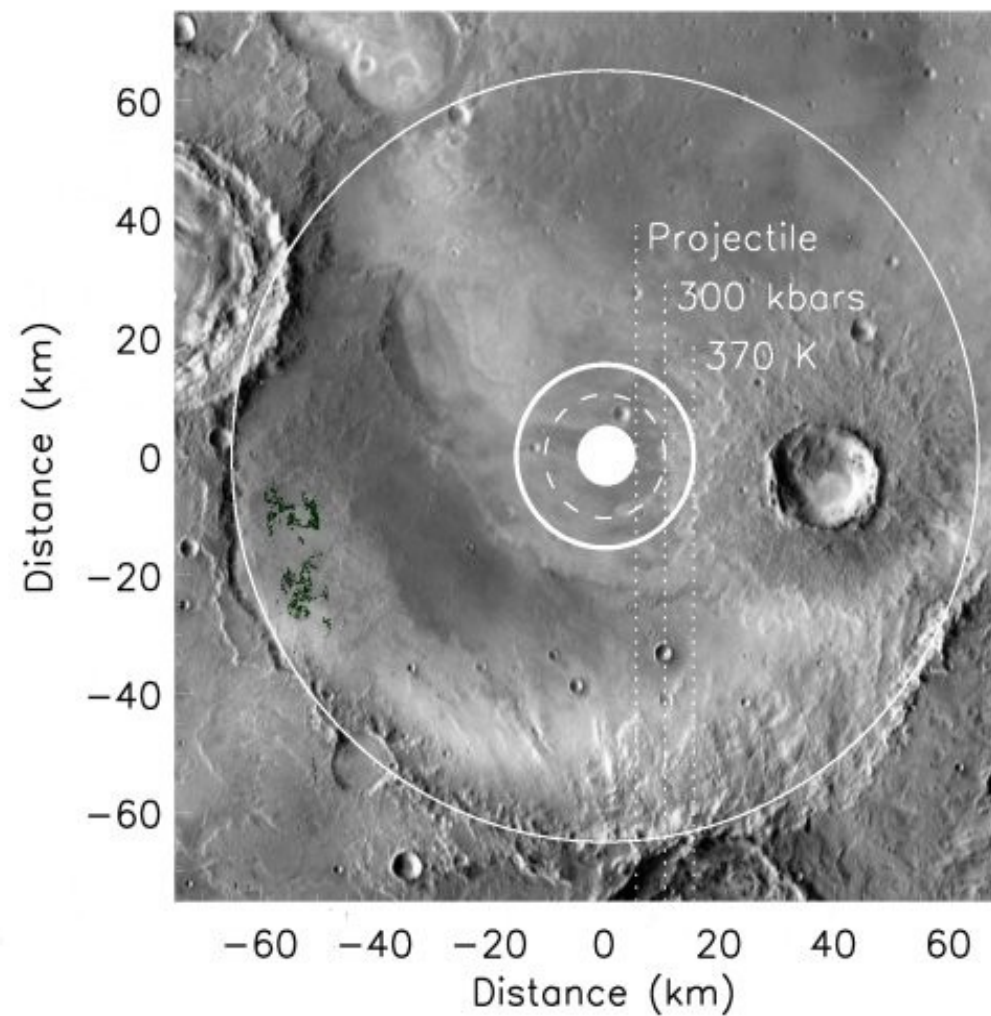
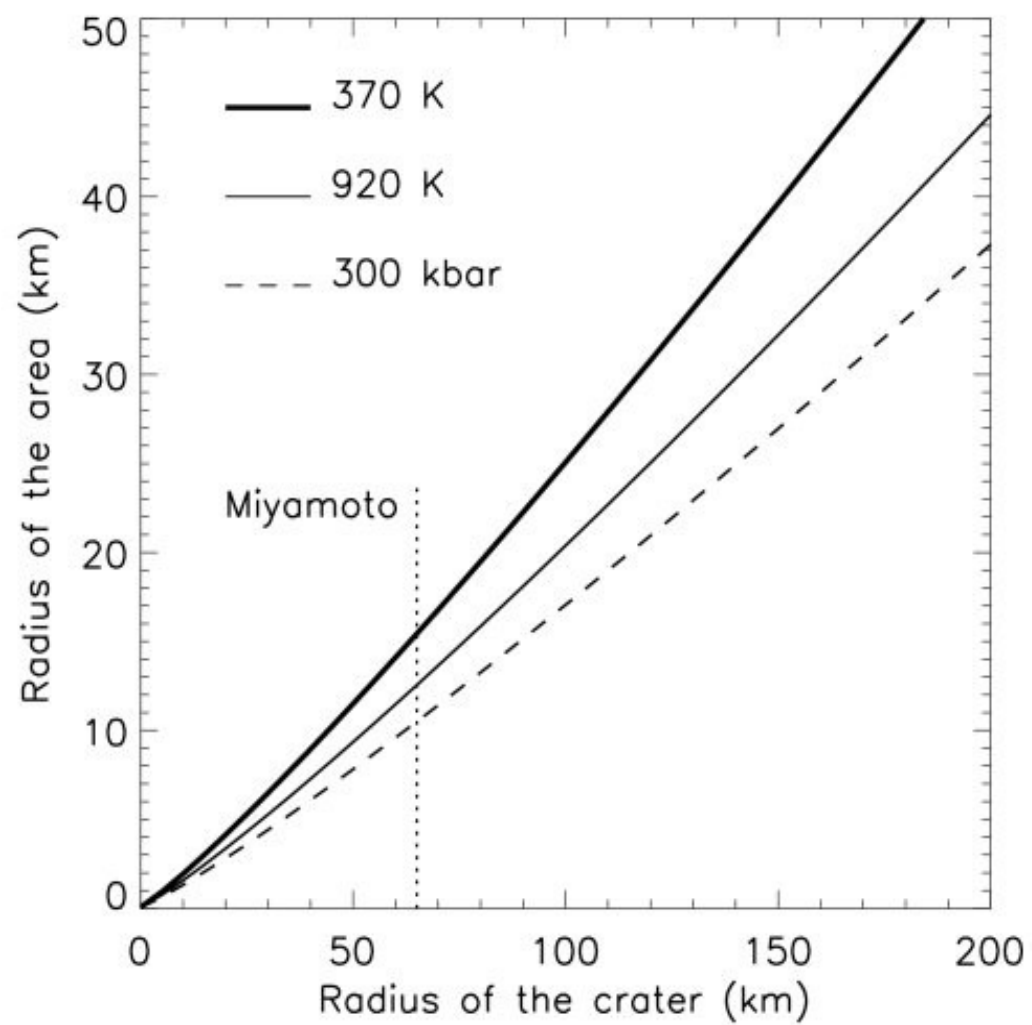
$$r_1 = r_0 \left( \frac{P_1}{P} \right)^{-\frac{1}{3.6}}$$

# The Model

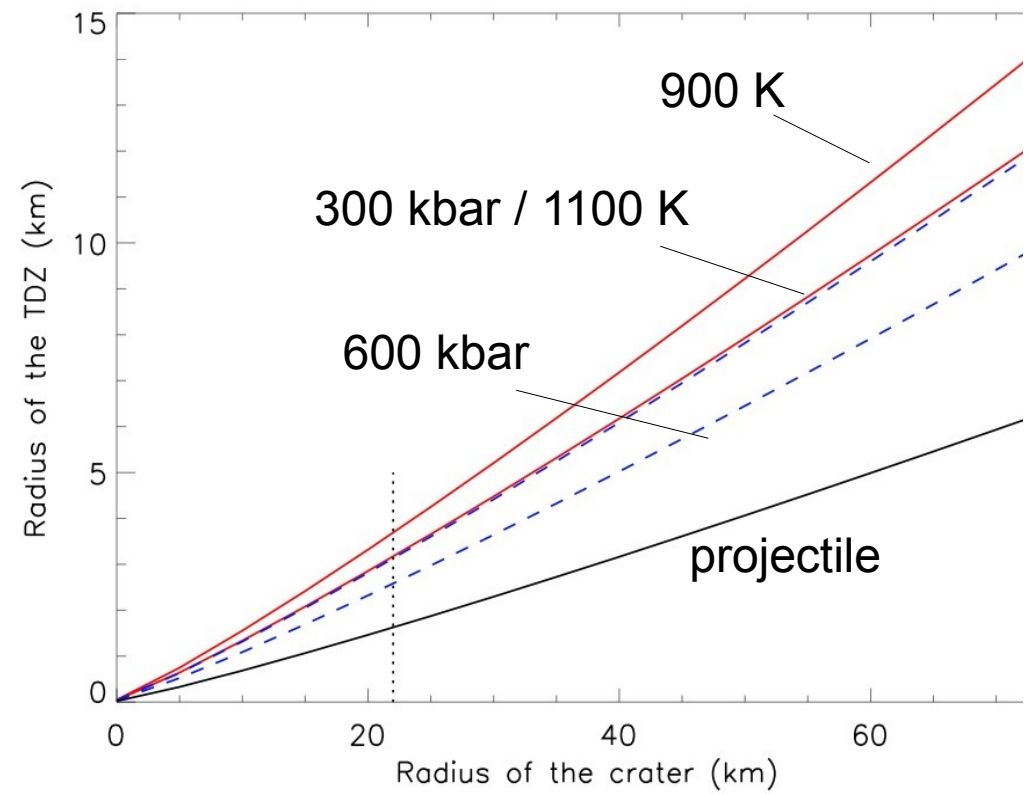
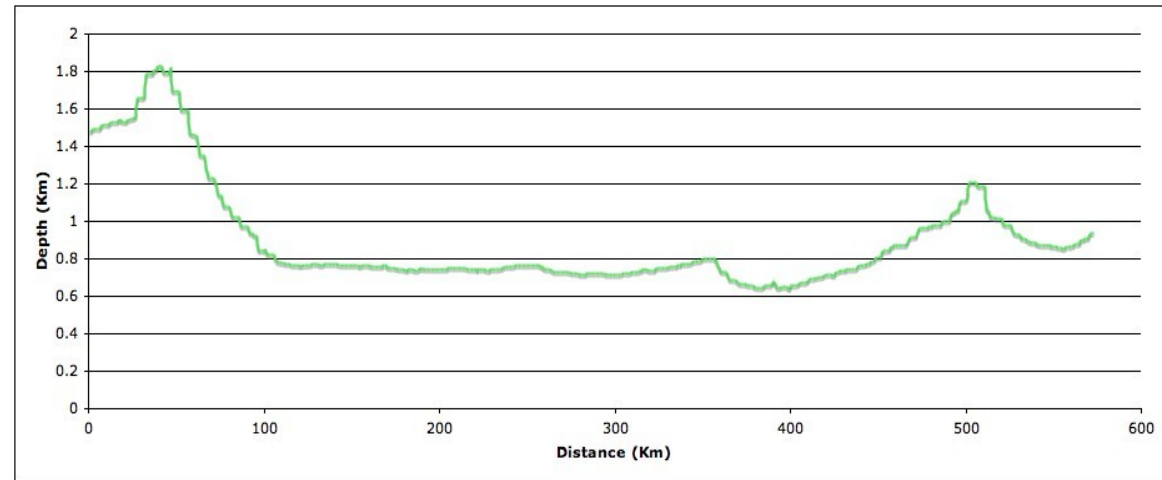
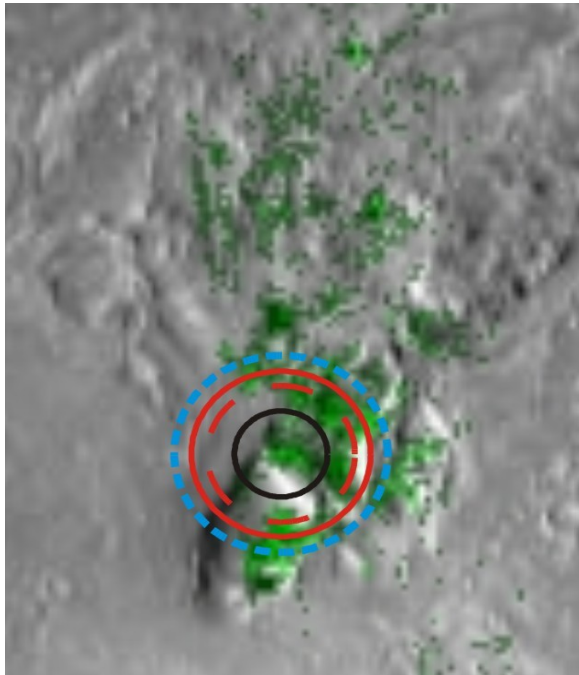
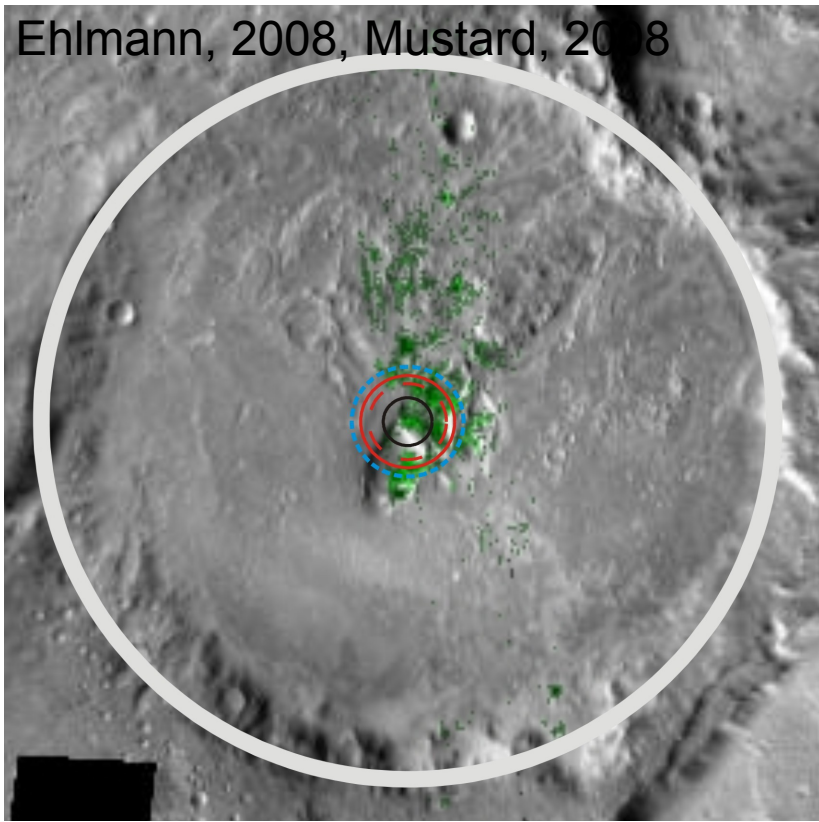
- For a basaltic target, the shock pressure required to heat the soil layer at a given depth to a temperature  $T$  can be estimated from Carr (1989):

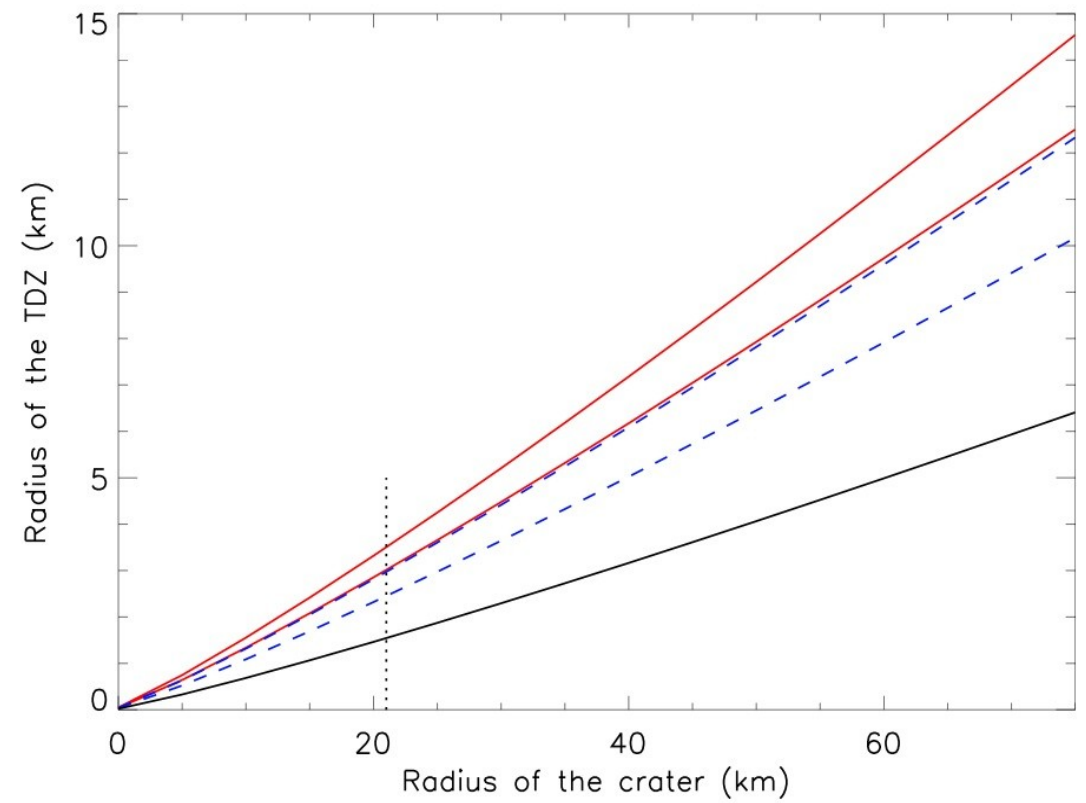
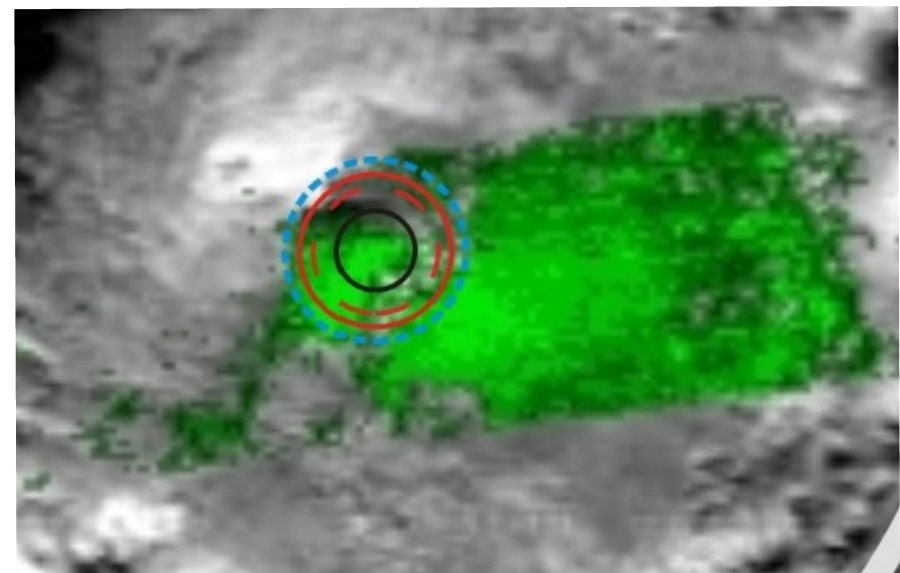
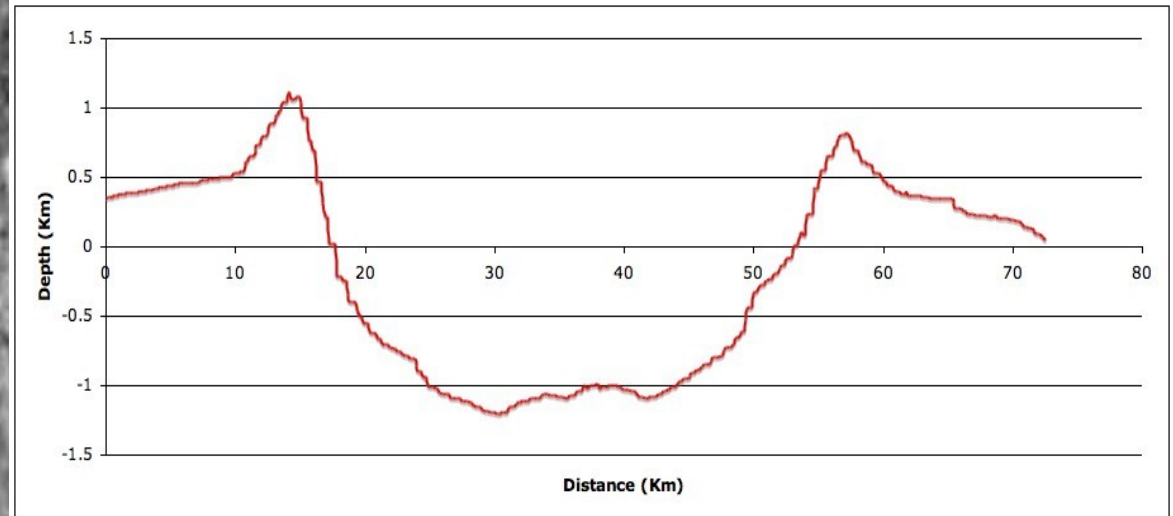
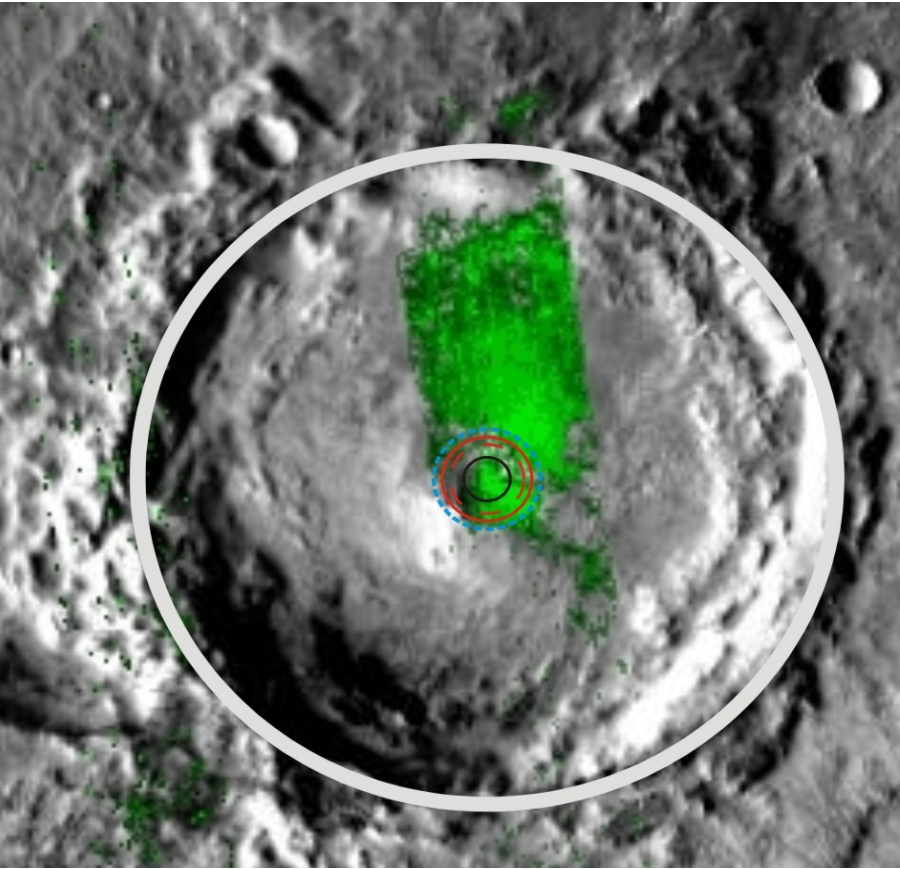
$$P = 3.279 + 0.9135(T - T_0) - 2.142 \times 10^{-3}(T - T_0)^2 + 1.748 \times 10^{-6}(T - T_0)^3$$



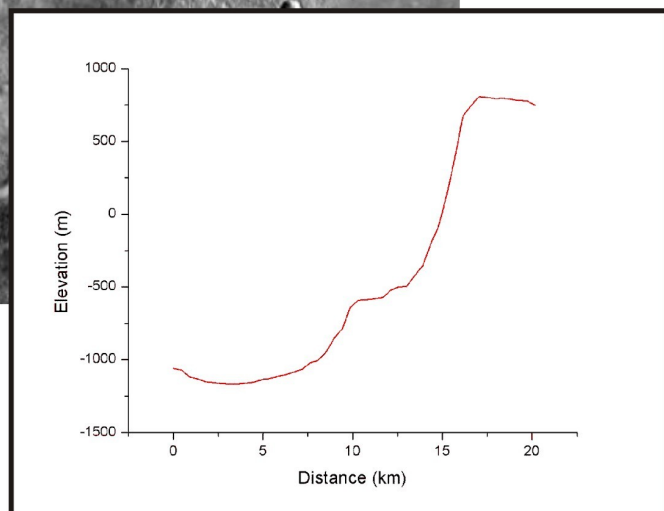
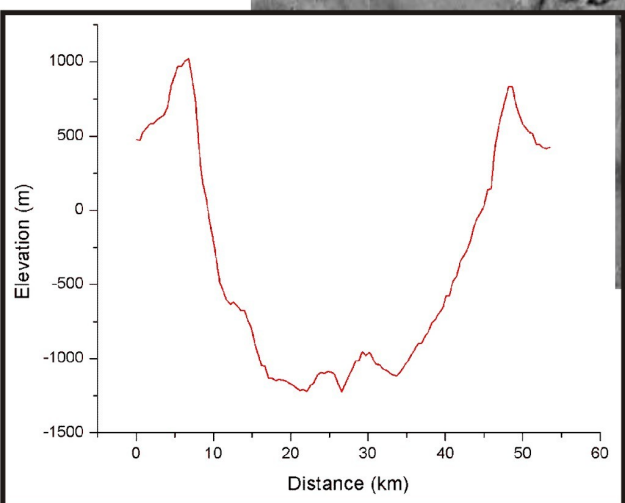
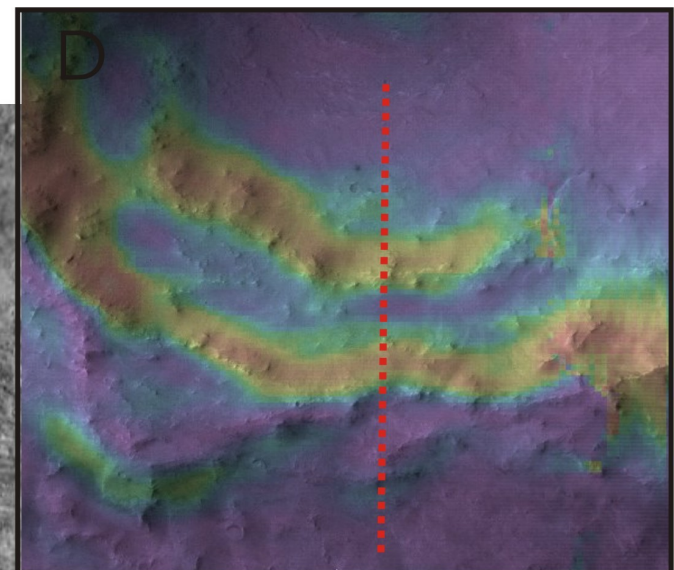
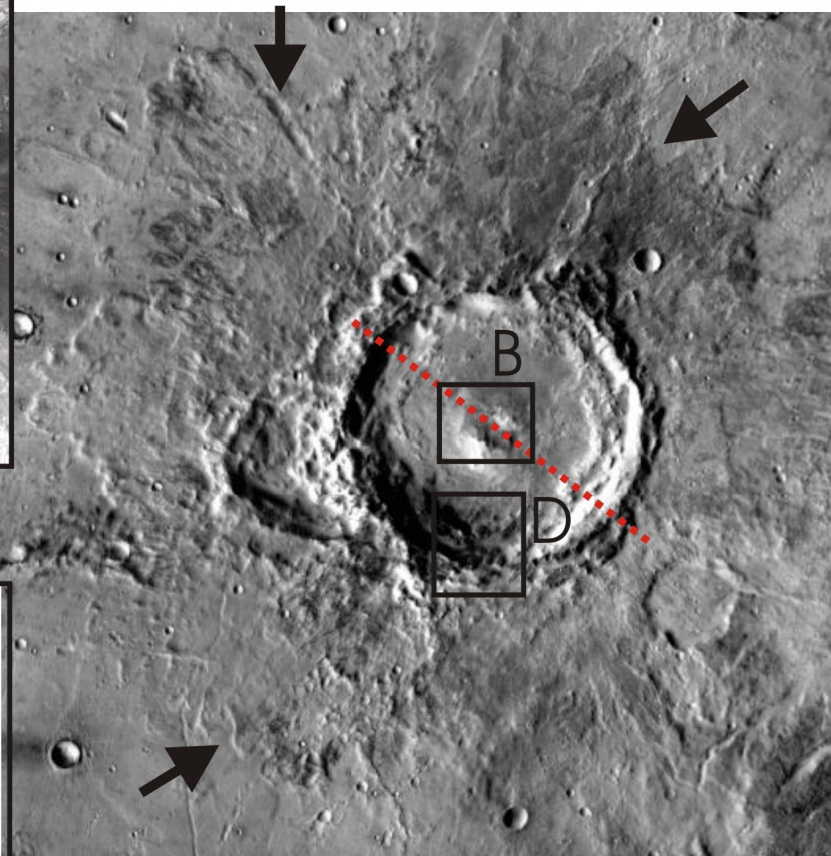
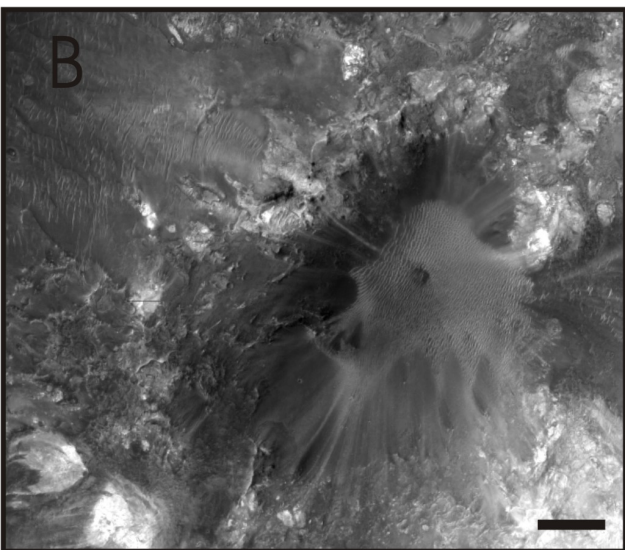


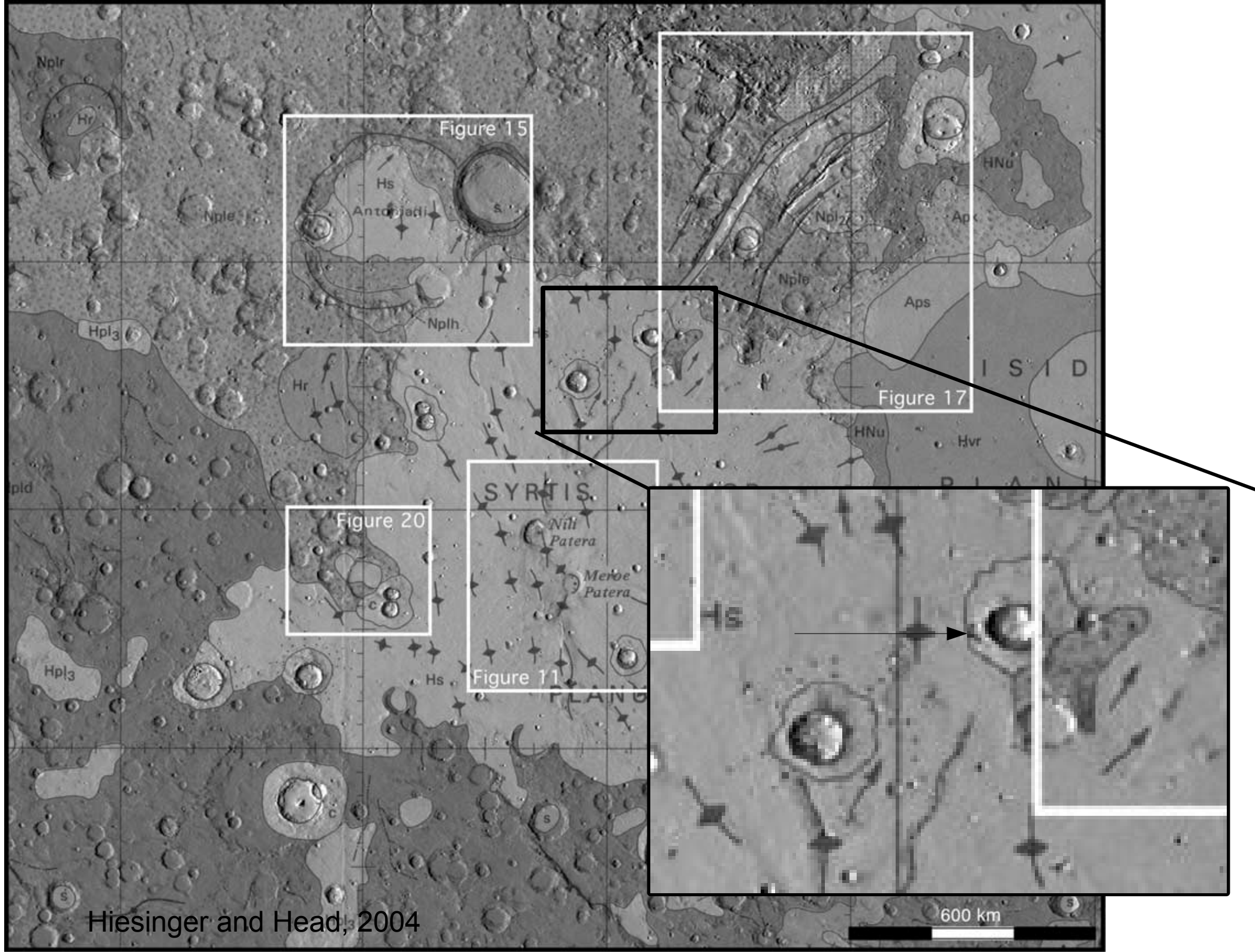
Ehlmann, 2008, Mustard, 2008



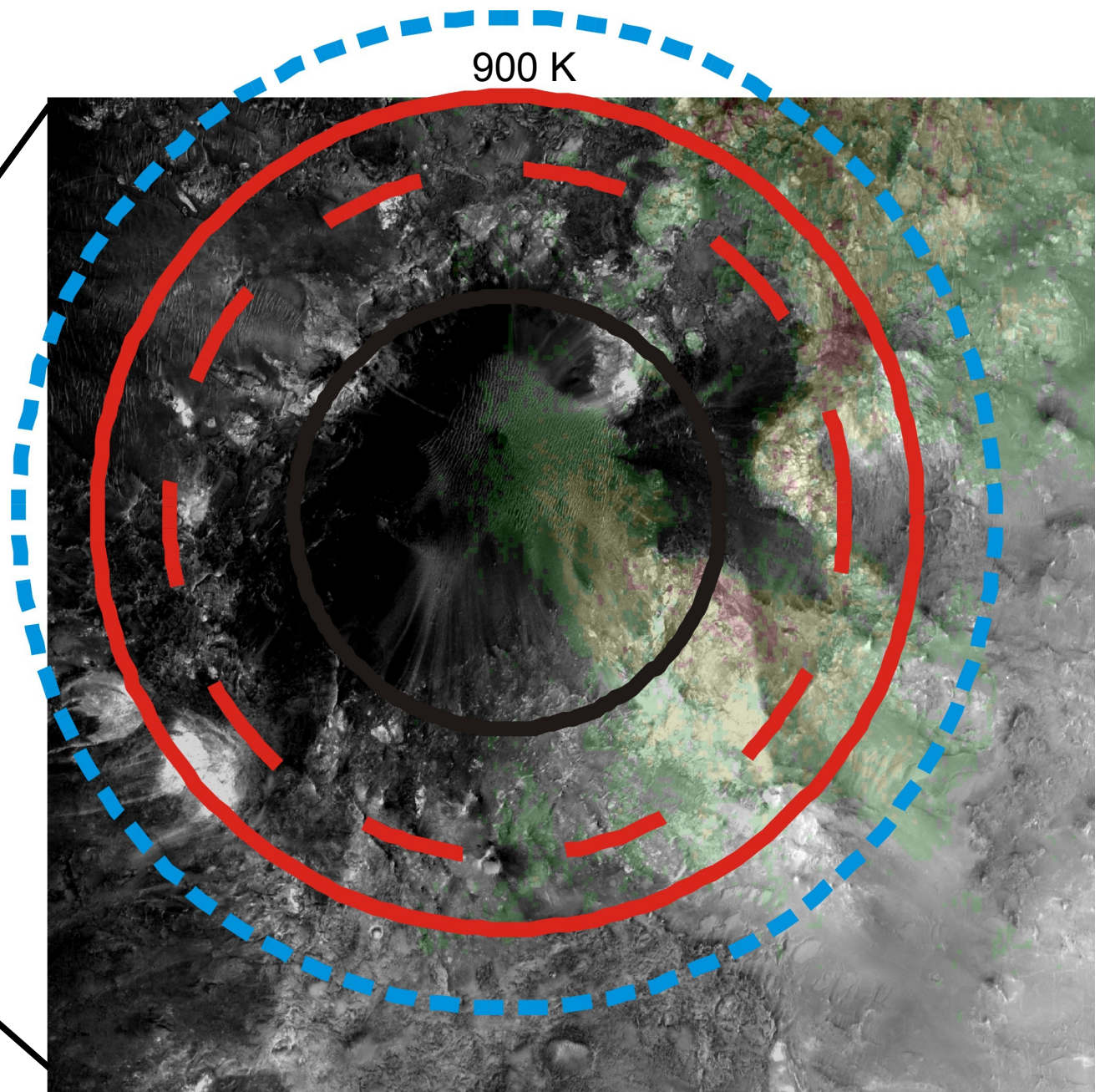
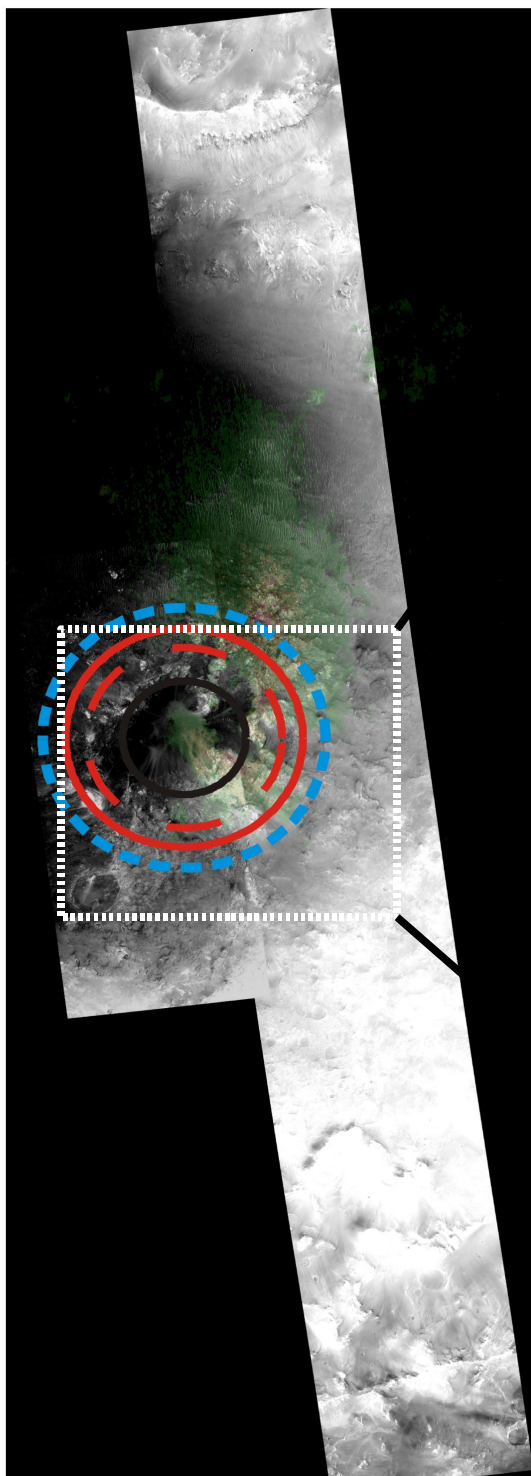








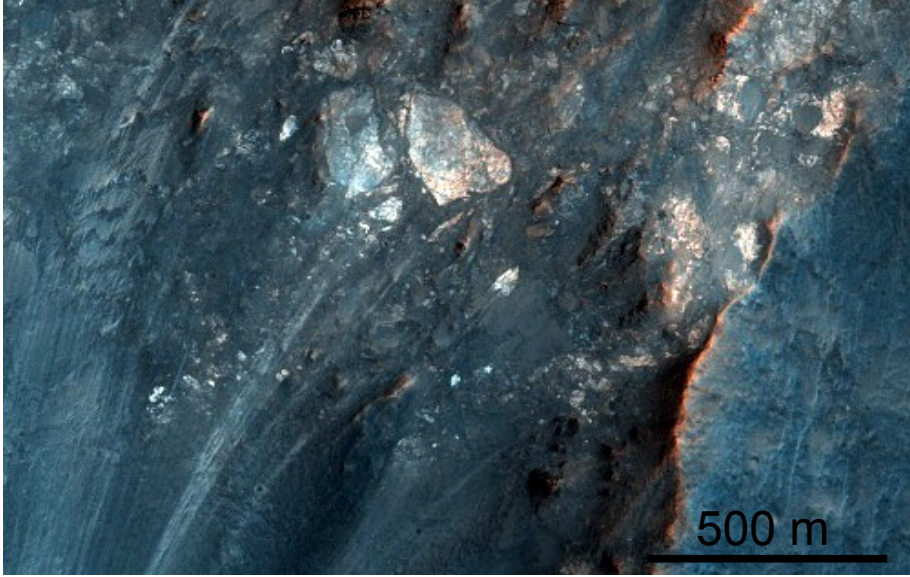




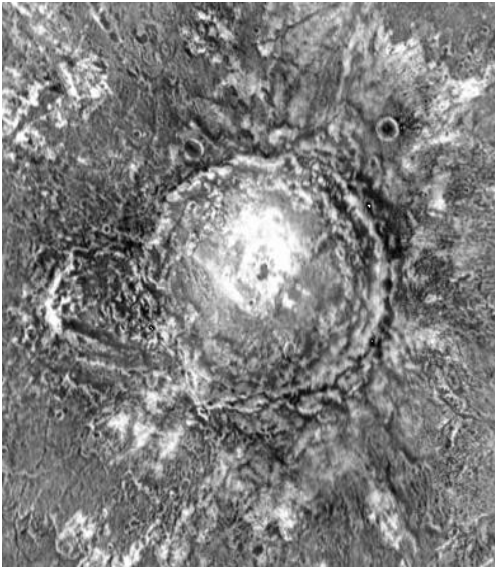
D2300 distribution from CRISM data  
Mg-Smectites (Nontronite? Saponite?) / Minor Montmorillonite?



# Eolian Transport?



NASA/JPL/ASU



Nighttime IR / THEMIS

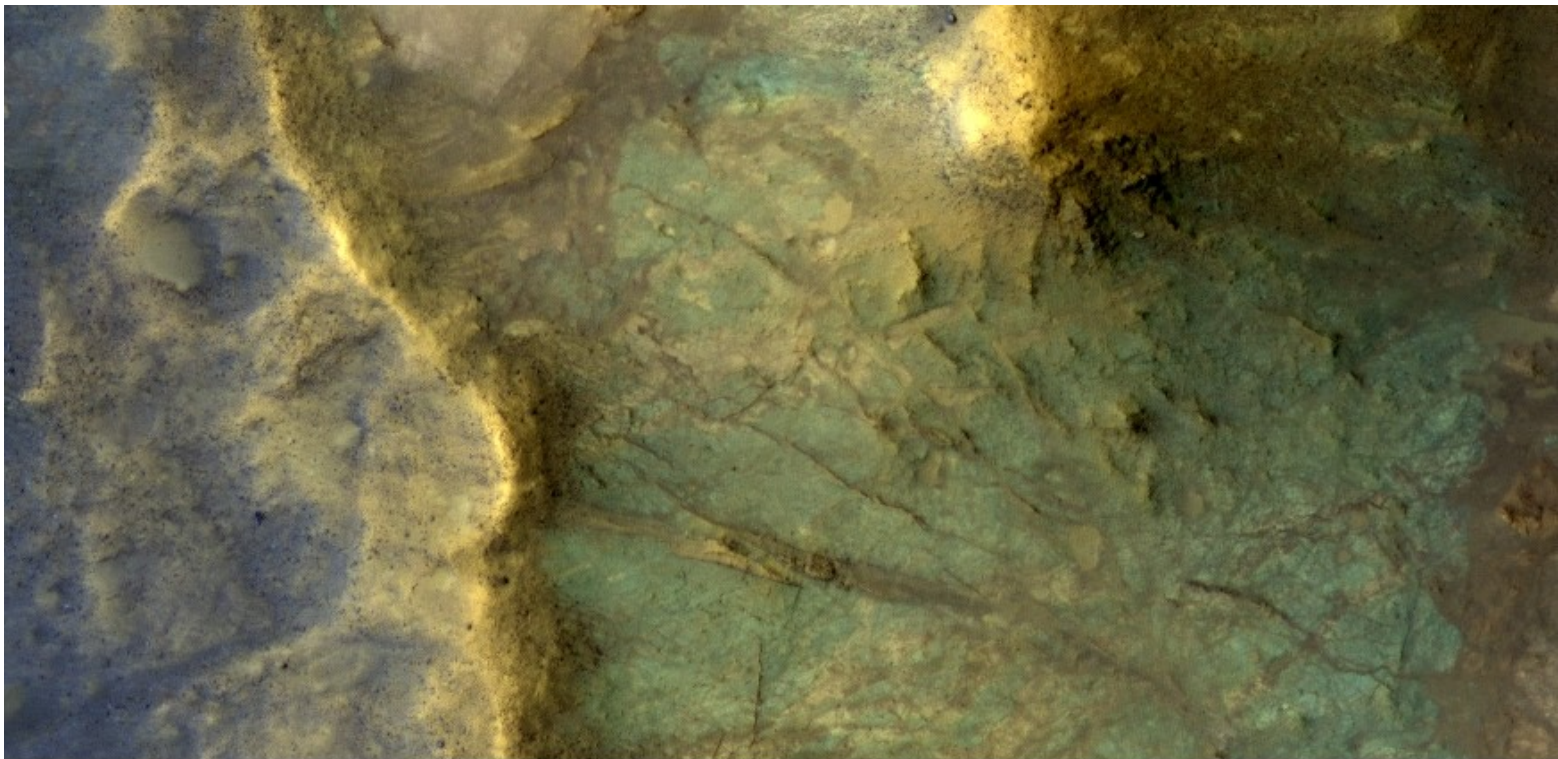
- Deposits have pristine appearance
- Discrete units up to several km size
- High Thermal Inertia (very coarse, indurated, or dust-free materials)
- Deposits occur on the crater floor, the highest parts of the central peak and along areas of the crater walls

# Fluvial Transport?

- Features associated with fluvial activity could not be identified in or around the crater at existing image resolution.
- Scenario remains a distinct possibility.

# Impact-induces hydrothermal activity

- Mineralogy of phyllosilicates within the TDZ compatible with hydrothermal origin.
- Impact-triggered hydrothermal activity can last over  $10^5$  yrs for a 40 km diameter crater (Rathbun & Squyres, 2002)
- Impact occurred in a volcanically active region (Syrtis Major)



# Summary

- We have presented a new tool to analyze the origin of phyllosilicates within impact craters (can be applied to other minerals too).
- Our model defines a Total Destruction Zone (TDZ), within which phyllosilicates would not have survived the impact event.
- Phyllosilicates within the TDZ of an impact crater could not have been excavated and their emplacement must post-date the impact event.
- A 40 km impact crater in Syrtis Major (> Hesperian), shows diverse phyllosilicates within the TDZ, which are better explained by aqueous processes (water transport or impact-triggered hydrothermal activity)

TABLE 7.1(a). Summary of thermal reactions – dioctahedral minerals

Kaolinite $\text{Al}_2\text{Si}_2\text{O}_5(\text{OH})_4$ 1:1 type	Predehydroxylation state → 450–550 °C Metakaolin $\text{Al}_2\text{Si}_2\text{O}_7$	→ 900 °C Spinel <sup>†</sup> -type phase	→ 1000–1100 °C Mullite <sup>†</sup> $\text{Al}_6\text{Si}_2\text{O}_{13}$
Pyrophyllite $\text{Al}_2\text{Si}_4\text{O}_{10}(\text{OH})_2$ 2:1 type	→ 650 °C Pyrophyllite dehydroxylate $\text{Al}_2\text{Si}_4\text{O}_{11}$	→ 950–1000 °C	Mullite <sup>†</sup> $\text{Al}_6\text{Si}_2\text{O}_{13}$
Montmorillonite <sup>‡</sup> 2:1 type	→ 150–250 °C Montmorillonite anhydride	→ 700 °C Montmorillonite dehydroxylate	1000 °C → Spinel-type phase 1150 °C Mullite <sup>†</sup>
Muscovite 2:1 type	→ 750 °C Muscovite dehydroxylate	→ 1050 °C Spinel-type phase Sanidine	→ 1250 °C Corundum $\alpha\text{-Al}_2\text{O}_3$ Mullite <sup>†</sup> Other Phases

<sup>†</sup> With separation of silica and/or other phases.

<sup>‡</sup> Ideal formula:  $\text{M}_{0.33}^+\text{nH}_2\text{O} \cdot (\text{Al}_{1.67}\text{Mg}_{0.33})\text{Si}_4\text{O}_{10}(\text{OH})_2$ .

Temperatures variable according to chemical composition, crystal size, conditions of heating.



TABLE 7.1(b) Summary of thermal reactions – trioctahedral minerals

Serpentine $\text{Mg}_3\text{Si}_2\text{O}_5(\text{OH})_4$ 1:1 type	→ 550 °C Serpentine dehydroxylate	→ 600 °C Forsterite (olivine) <sup>†</sup> $\text{Mg}_2\text{SiO}_4$	→ 1100 °C Forsterite + enstatite $\text{MgSiO}_3$
Talc $\text{Mg}_3\text{Si}_4\text{O}_{10}(\text{OH})_2$ 2:1 type	→ 900–1000 °C Enstatite <sup>†</sup>		
Saponite <sup>‡</sup> 2:1	→ 550 °C Saponite dehydroxylate	→ 750 °C Enstatite <sup>†</sup>	
Vermiculite <sup>§</sup> 2:1	→ 500–800 °C Vermiculite dehydroxylate	→ 850 °C Enstatite <sup>†</sup>	
Chlorite $(\text{Mg},\text{Al})_6(\text{Si},\text{Al})_4\text{O}_{10}(\text{OH})_8$ 2:1 type	→ 450–550 °C partial or complete dehydroxylation	→ 800 °C Olivine <sup>†</sup>	

<sup>†</sup> With separation of silica and/or other phases.

<sup>‡</sup> Ideal formula:  $\text{M}_{0.33}^+\text{nH}_2\text{O} \cdot \text{Mg}_3(\text{Si}_{3.67}\text{Al}_{0.33})\text{O}_{10}(\text{OH})_2$ .

<sup>§</sup> Typical formula:  $\text{Mg}_{0.38}^{2+}\text{nH}_2\text{O} \cdot (\text{Mg}_{2.00}\text{Fe}_{0.46}^{\text{III}}\text{Al}_{0.22}\text{Ti}_{0.11})(\text{Si}_{2.72}\text{Al}_{1.28})\text{O}_{10}(\text{OH})_2$ .

Temperatures variable according to chemical composition, crystal size, condition of heating.



