CONTROLS ON THE POTENTIAL OCCURRENCE OF GROWING, ERUPTIBLE MARTIAN CRUSTAL MAGMA CHAMBERS.

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Introduction: The depth at which magma pools in the crust exercises a key control over the extent of the physical and chemical evolution of the resulting magma chamber. On Earth, long-lived silicic magma chambers that can erupt tend to accumulate at 2±0.5kbar [1]. As with terrestrial magma chambers, the possibility of creating a long-lived eruptive magma chamber in the Martian crust would require crustal conditions that allow the chamber to grow by mass, being fed by magma recharges while simultaneously being conducive to the exsolution of a volatile phase, yet not allowing the chamber to cool too rapidly.

Previous estimates of magma storage pressure using mineral geobarometry from meteorites [2,3] suggest entrapment at varying depths within the Martian crust, but these results remain questionable and need to be tested by independent methods. Caldera subsidence in Olympus Mons suggests that magma reservoirs may be situated ≤16km depth [4], hinting the existence of shallow crustal magma reservoirs which could fuel episodic eruptions. Recently, the NASA InSight mission has reported the occurrence of several low-frequency (LF) marsquakes at around Cerberus Fossae (CF) and analysis of the spectral characters of these events indicates their origin to be around 30-50km, in a structurally weak and potentially warm source region [5]. It has also been proposed that these LF events could originate from a low-viscosity, high-flux fluid flow [6], rendering the eruptive nature of these hypothetically active magma reservoirs ambiguous.

In general, Martian crustal magma reservoirs have been proposed to be situated deeper than their counterparts on Earth [7] because of lower gravity, colder conditions (at least since the last 1 Ga), and lower H2O/CO2 ratios [8,9], but it has not been quantified. Using thermo-mechanical modelling, we aim to study magma chamber evolution at different levels in the Martian crust & whether a range of conditions including magma chamber size, magma recharge rate, volatile content exists such that magma chambers grow while being simultaneously susceptible to eruptions.

Method: We have used the thermo-mechanical magma chamber model of [10,11] to simulate the evolution of magma bodies in the Martian crust;

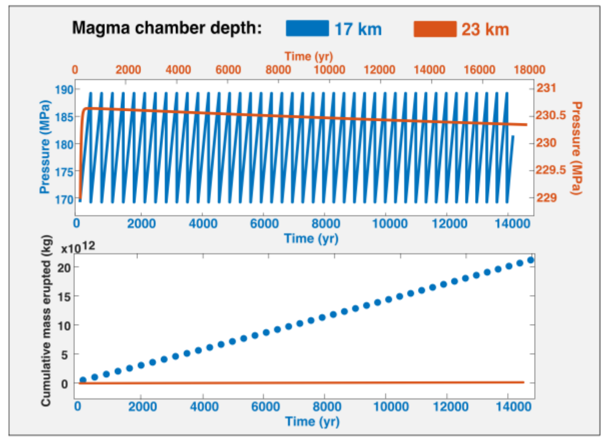

Figure 1: A comparative time-series output for two simulated Martian crustal magma chambers at 17km and 23km depths, showing the evolution of pressure (above) and cumulative erupted mass (below) from the chamber. Both chambers have identical volumes (50km^3), inflow rates (10^{-3}km^{3}yr^{-1}) and volatile contents (0.1wt% H_2O, 100ppm CO_2). The 17km chamber (blue) shows a higher eruption frequency (each pressure spike indicates an eruption), expelling a significant amount of mass, but is comparatively more short-lived than the chamber at 23km depth (orange), which being surrounded by a warmer, more viscous crust, does not erupt and also experiences slower cooling.

wherein the magma chamber experiences a constant recharge while losing heat by conduction to the surrounding visco-elastic shell, causing crystallization and volatile exsolution. Chamber behaviour (growth, eruption frequency and crystallization) for different starting conditions can be cast in the interplay of three competitive timescales; injection(τ_{in}), cooling(τ_{cool}) & viscous relaxation(τ_{relax}). The model was run at various simulated depths (1-2.5kbar) of the Late Amazonian (current) Martian crust to constrain the conditions that satisfy magma chamber growth and eruptions for a range of chamber sizes, long-term average recharge rates and magma volatile contents (Fig. 1). Based on observations, the simulations were carried out in two parts, one for the overall Martian crust and the other focusing on Cerberus Fossae.

Results: Some of the main observations we gather from the simulations are as follows:

General Martian crust. At high pressures of ≥2.5kbar (≥25km), obtaining a simultaneously eruptive and growing magma chamber is challenging, irrespective of initial chamber volumes and recharge rates. Largest accumulation of mass (coupled with volume growth) is seen in the smallest chambers with the highest influx. However, these chambers are least susceptible to producing eruptions, being hosted by a viscous, accommodating crust.

At pressures < ~1.5kbar (≤17km), minimal growth (≤10%) is observed in simulations where τ_{cool} exceeds τ_{in}, irrespective of initial volume. For chambers where τ_{in} exceeds τ_{cool}, the lifetime of the chamber is short, which prevents significant growth.

At depths of 20-23km, a narrow set of conditions exists for chambers where significant growth occurs along with eruptions. If the viscous deformation of the crust is efficient and prevents pressure build-up ($\tau_{relax}<\tau_{in}$), the chamber grows but does not erupt. In contrast, when $\tau_{relax}>\tau_{in}$, the injection rate outpaces pressure relaxation, resulting in a high eruption frequency but lower chamber growth and longevity. If the chamber cools efficiently ($\tau_{cool}<\tau_{in}$) and crystallization outpaces injection, the chamber does not grow similarly to its shallower counterparts. Hence, simultaneous growth and eruption would only be possible when τ_{relax} and τ_{in} are comparable and τ_{cool} exceeds τ_{in} (injection outpaces cooling).

Cerberus Fossae (CF): The postulated depth of origin of the LF seismic events have a large range (15-50km) [12]. Based on the nature of the LF events, they have been suggested to originate in a warm, ductile source region (assumed to be 40±10km in [5]), which suggests a locally higher thermal gradient of 20±2K/km (compared to 15K/km [13] used for the general cases). Our results indicate that under these conditions and over the range of studied recharge rates (10^{-5} to 10^{-3} km^3/yr), we find that it is impossible for a magma chamber situated deeper than 20km to contribute to the surface volcanic record. In this case, the optimal depth of growth and eruption would be pushed upwards ~15km. Young ages of surface volcanics (Cerberus Fossae Mantling Unit; CFMU) have been reported from CF (53ka-210ka) [14]. However, given that the relatively fast cooling at these shallow depths would make it challenging to keep such an eruptible chamber active for at least ~50ka, it suggests the episodic nature of the magma source region of CFMU. Nevertheless, the non-erupting 40km chamber could potentially remain active for ≥60ka with these inflow rates if its volume is at least in the order of 10^2 km^3.

Conclusions: Based on ~5000 thermo-mechanical simulations of mafic magma reservoirs in the Late-Amazonian Martian crust, we stress that magma chambers which can simultaneously erupt and grow could only potentially exist around 20km depth

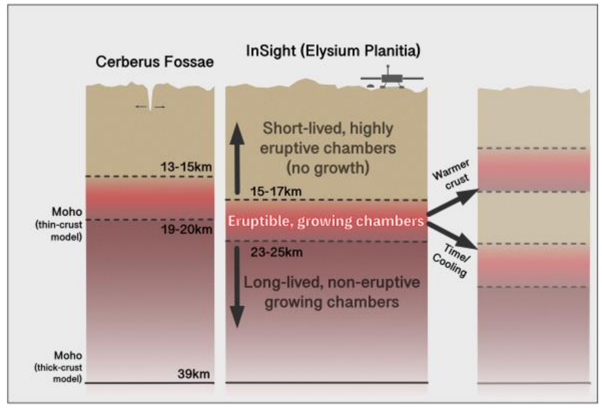

Figure 2: A schematic cross-section of the crust under Elysium Planitia showing the three regimes of magma chamber behaviour. The zone of growing + erupting chambers shows a shallowing trend for warmer crust (e.g., under Cerberus Fossae), while a colder, more elastic crust would push it deeper. Crustal thickness data from [15].

(2kbar); while this depth is affected by the thermal gradient (Fig 2), their existence is dependent on recharge rate and chamber volume. A corollary to this conclusion is that the formation of evolved "granitic" compositions in the shallow Late-Amazonian (<1 Ga) magma chambers would be hindered by fast cooling and frequent eruptions.

References: [1] Huber et al. (2019) *Nature Geoscience*, **12(9)**, 762-768. [2] Rahib et al. (2019), *GCA*, 463-496. [3] Filiberto et al. (2010) *Meteoritics & Planet. Sci.*, **45(8)**, 1258-1270. [4] Zuber and Mouginis-Mark (1992), *JGR: Planets*, **97(E11)**, 18295-18307. [5] Stähler et al. (2022), *Nature Astronomy*, 1-11. [6] Kedar et al. (2021), *JGR: Planets*, 126, e2020JE006518. [7] Wilson and Head, (1994), *Reviews of Geophysics*, **32(3)**, 221-263. [8] Black et al. (2022), *GRL*, **49(11)**, e2022GL098061. [9] Stanley et al. (2011), *GCA*, **75(20)**, 5987-6003. [10] Degruyter and Huber, (2014), *EPSL*, 403, 117-130. [11] Townsend et al. (2019), *Geology, Geophysics, Geosystems*, **20(3)**, 1574-1591. [12] Brinkman et al. (2021), *JGR: Planets*, **126(4)**, e2020JE006546. [13] Degruyter et al. (2017) *LPS XLVIII*, Abstract #2034. [14] Horvath et al. (2021), *Icarus*, 365, 114499. [15] Knapmeyer-Endrun et al. (2021), *Science*, **373(6553)**, 438-443.