



**HAL**  
open science

# Magma ascent at floor-fractured craters diagnoses the lithospheric stress state on the Moon

Chloé Michaut, Virginie Pinel, Francesco Maccaferri

## ► To cite this version:

Chloé Michaut, Virginie Pinel, Francesco Maccaferri. Magma ascent at floor-fractured craters diagnoses the lithospheric stress state on the Moon. *Earth and Planetary Science Letters*, 2020, 530, pp.115889 -. 10.1016/j.epsl.2019.115889 . hal-03488414

**HAL Id: hal-03488414**

**<https://hal.science/hal-03488414v1>**

Submitted on 21 Dec 2021

**HAL** is a multi-disciplinary open access archive for the deposit and dissemination of scientific research documents, whether they are published or not. The documents may come from teaching and research institutions in France or abroad, or from public or private research centers.

L'archive ouverte pluridisciplinaire **HAL**, est destinée au dépôt et à la diffusion de documents scientifiques de niveau recherche, publiés ou non, émanant des établissements d'enseignement et de recherche français ou étrangers, des laboratoires publics ou privés.



Distributed under a Creative Commons Attribution - NonCommercial 4.0 International License

# Magma ascent at floor-fractured craters diagnoses the lithospheric stress state on the Moon

Chloé Michaut

*Laboratoire de Géologie de Lyon, Université de Lyon, Ecole Normale Supérieure de Lyon,  
Université Lyon-1, CNRS, UMR 5276, 46 allée d'Italie, F-69364, Lyon, France*

Virginie Pinel

*Université Grenoble Alpes, Univ. Savoie Mont Blanc, CNRS, IRD, IFSTTAR, ISTERre, 38000  
Grenoble, France*

Francesco Maccaferri

*German Research Centre for Geoscience (GFZ), Telegrafenberg, 14473, Potsdam, Germany and  
Istituto Nazionale di Geofisica e Vulcanologia INGV, Osservatorio Vesuviano, Napoli, Italy*

---

## Abstract

On the Moon, floor-fractured craters (FFCs) present evidence of horizontal crater-centred magmatic intrusions. Crater floor uplift and moat formation indicate that these sill intrusions occur at shallow depths ( $< 10$  km). While a recent study has demonstrated that magma ascent below FFCs and mare-filled craters was triggered by crater unloading, the mechanism leading to the emplacement of shallow sills is still poorly understood. Here we show that the local stress field due to crater unloading is also responsible for the horizontalisation of the magma flow leading to sill-like intrusions. On Earth, caldera formation has been shown to similarly affect magma trajectories, inducing the formation of a sill-shaped storage zone. Magma ascent to shallow depths below FFCs was however made possible because of a regional tensional stress caused by mare loading on the lunar lithosphere. We show that the tensional stress generated by elastic lithosphere deformation

*Preprint submitted to Earth Planet. Sc. Lett.*

*March 8, 2020*

caused by mare loading combined to the local crater stress field can explain the distribution of FFCs on the Moon, with the smallest FFCs being located over a larger distance range from the mare. In particular, FFCs distribution around Oceanus Procellarum is consistent with an average load thickness of  $\sim 1$  km. This study suggests that magma trajectory in the crust of terrestrial planets can provide a diagnostic of the lithospheric structure and state of stress.

*Keywords:*

magma intrusion, crustal stress field, magma path, floor-fractured crater, lunar mare

---

## 1. Introduction

2 In the crust of terrestrial planets, magma transport occurs by magma flow  
3 through induced fractures referred to as dykes or sills depending on their ori-  
4 entation relative to lithologic layers or verticality. Observations show that hori-  
5 zontal sills and laccoliths on Earth are fed by vertical dykes and inversely, dykes  
6 are fed by horizontal sills in large sill complexes. (Cartwright and Moller Hansen,  
7 2006; Muirhead et al., 2012; Richardson et al., 2015): intrusions thus deviate from  
8 their initial orientations. The presence of inherited structures or the reaching of a  
9 neutral buoyancy zone are often advocated to explain deviations in magma paths  
10 (Menand, 2011). Recently, studies on the trajectory of magma in the Earth's crust  
11 have shed light on the crucial influence of the stress field on magma trajectory and  
12 its ability to reach the surface (Pinel et al., 2017; Corbi et al., 2015; Maccaferri  
13 et al., 2014). Below a volcanic edifice, the induced compressive stress opposes  
14 to magma ascent, allowing only for the ascent of the least dense and thus most  
15 differentiated magmas (Pinel and Jaupart, 2000). Providing the magma driving

16 force (overpressure) is weak, the crustal stress field may exert a strong control  
17 on the trajectories of magma. Magmatic dykes tend to open in the direction of  
18 least compressive stress ( $\sigma_3$ ) (Anderson, 1951; Nakamura, 1977). Analog experi-  
19 ments (Watanabe et al., 2002; Corbi et al., 2016) and numerical models (Mériaux  
20 and Lister, 2002; Maccaferri et al., 2011) indeed show that dyke paths are more  
21 influenced by the local stress if the magma driving force is weak.

22 On the Moon, floor-fractured craters (FFCs) are impact craters that were mod-  
23 ified after their formation. Compared to non-modified craters, they show fractured  
24 and uplifted floors that are believed to result from crater-centred horizontal sill or  
25 laccolith intrusions (Schultz, 1976a; Wichman and Schultz, 1995; Jozwiak et al.,  
26 2012; Thorey and Michaut, 2014). The magma feeding these intrusions must have  
27 been transported through vertical dykes and then turned into a horizontal sill at a  
28 shallow depth to explain the observed floor uplift (Jozwiak et al., 2015; Wilson  
29 and Head, 2017). Deviation from the vertical has previously been explained by  
30 the presence of a highly brecciated zone caused by the impact. Schultz (1976a)  
31 proposed that the magma reaching the breccia lens filled the subhorizontal inher-  
32 ited fractures forming a sill while the impact melt sheets welded the crater floor,  
33 leading to cohesive floor deformation and uplift. Jozwiak et al. (2012, 2015) al-  
34 ternatively proposed that the vertical ascent of magma was halted because of the  
35 low density of the brecciated zone. The impact crater topography naturally results  
36 from crustal material redistribution, which can be described by surface unload-  
37 ing forces. The elastic stresses underneath impact craters have been studied by  
38 Melosh (1976) but the effect of unloading forces on the principal stresses direc-  
39 tion beneath a crater has never been considered in the context of FFC formation.  
40 On Earth, it has been shown that unloading forces due to rift valleys and calderas

41 formation would affect magma pathways (Maccaferri et al., 2014; Corbi et al.,  
42 2015). In fact, below a surface unloading, the direction of  $\sigma_3$  is often vertical,  
43 favouring the formation of horizontal sills (Corbi et al., 2015; Pinel et al., 2017)  
44 in case the magma driving force is weak.

45 The lunar Highlands, where many FFCs are present, are composed of ancient,  
46 highly porous, anorthositic rocks (Wood et al., 1970). Analysis of GRAIL's grav-  
47 ity data show that this ancient primary crust has a low average density of  $\sim 2550$   
48  $\text{kg m}^{-3}$  (Wieczorek et al., 2013). The buoyancy of basaltic magma in such a low  
49 density crust is very limited, therefore, the resulting magma overpressure may be  
50 small enough for magma trajectories to be sensitive to the local stress field at the  
51 time of intrusion.

52 Michaut and Pinel (2018) have recently shown that crater unloading can trig-  
53 ger magma ascent in the lunar crust, the decompression of the encasing rocks  
54 providing a relative overpressure to the magma that was ponding at depth. Obser-  
55 vations indeed show that for large craters on a thin crust, magma erupted on the  
56 floor of mare-filled craters and of some FFCs, suggesting that the overpressure  
57 provided by the unloading was large enough for the magma to reach the surface.  
58 However, for smaller craters or a thicker crust, crater floor uplift is observed sug-  
59 gesting that the overpressure was too small for the magma to reach the surface  
60 (Michaut and Pinel, 2018). Here we show that, below such craters, the local stress  
61 field due to crater unloading is the cause of magma path deviation from a vertical  
62 dyke to a horizontal sill. This sill formation is however expected to occur close  
63 to the crust base, preventing any further ascent. It follows that the emplacement  
64 of shallow intrusions cannot be explained considering only the crater unloading  
65 effect. We demonstrate that the presence of dense basaltic maria and mascons in

Table 1: Defining characteristics of each classes of floor-fractured crater, after Schultz (1976a) and Jozwiak et al. (2012).

Crater Class	Characteristics
1	Deep, fractured floors, mare material along walls
2	Well defined wall scarp, uplifted central region / convex up floor profile, concentric fractures
3	Wide U-moat between wall scarp and uplifted crater interior, radial and polygonal fractures
4 a,b,c	V-shaped moat with a convex up floor profile to hummocky interior
5	Degraded walls and fractured floors
6	Mare-flooded interiors, concentric fracture near wall

66 the nearby region added an important contribution to the lithospheric stress field,  
 67 allowing for a vertical magma ascent up to shallow depths. By taking into account  
 68 maria-induced stress, we explain the size and spatial distribution of FFCs around  
 69 the lunar maria and use it to constrain mare thicknesses.

## 70 **2. Observations on lunar floor-fractured craters**

71 FFCs are craters presenting radial and concentric floor fractures. Their floors  
 72 are generally uplifted compared to non-modified craters of the same size. Schultz  
 73 (1976a), and lately Jozwiak et al. (2012), proposed a classification into six differ-  
 74 ent FFC classes on the Moon based on crater floor morphologies and appearances  
 75 (Table 1). Crater-centred intrusion of basaltic magma is believed to be the cause

76 of floor deformation and uplift (Schultz, 1976a; Jozwiak et al., 2012; Thorey et al.,  
77 2015). In fact, volcanic deposits are often present within these craters, and theoret-  
78 ical models for the dynamics of crater-centred horizontal intrusions well explains  
79 the diversity of crater floor appearances (Thorey and Michaut, 2014). Further-  
80 more, FFCs are distributed around basaltic maria and mascons (Figure 1), at dis-  
81 tances reaching one to two thousands of kilometers from these maria. Craters of  
82 Class 4, although smaller in size than those of other classes, occur over a greater  
83 distance range from the maria, deeper into the Highlands (green dots on Figure 1)  
84 (Schultz, 1976a).

85 Craters of Classes 2, 3 and 4 present an uplifted floor (dots on Figure 1), that  
86 appears either convex up or flat with a circular moat bordering the inner side of  
87 the crater wall (Table 1). Such floor appearances are characteristics of large sill-  
88 like intrusions (with opening of up to several hundred meters to a few kilometers)  
89 at shallow depth (Thorey and Michaut, 2014) and hence of a strong deviation of  
90 the magma paths from vertical feeding dikes to horizontal sills occurring lately in  
91 the ascent. For such craters, the effective overpressure provided to the magma by  
92 the crater unloading was not large enough to allow for eruption and magma got  
93 trapped at depth (Michaut and Pinel, 2018). Thus, the magma driving pressure  
94 was supposedly small and the magma was likely sensitive to the lithospheric stress  
95 state (Watanabe et al., 2002).

96 Craters of Class 1 show fractures and mare patches along the crater wall but  
97 their floors have not been uplifted. Such characteristics do not point to the pres-  
98 ence of a large sill-like intrusion. Craters of Class 6 are completely flooded and  
99 it is thus not possible to know if a large-volume horizontal intrusion is present at  
100 depth. Craters of Class 1 and 6 are generally large craters (Jozwiak et al., 2012)

101 on a relatively thin crust such that the crater unloading generally provided enough  
102 effective overpressure for the magma to reach the surface (see Figure 3 of Michaut  
103 and Pinel (2018)). For those two classes, the driving pressure was probably too  
104 large for the magma to be very sensitive to the lithospheric state of stress (Michaut  
105 and Pinel, 2018).

106 Craters of class 5 are old and degraded (Jozwiak et al., 2012) and are not  
107 considered in our analysis.

108 We assume that the initial vertical ascent of the magma was triggered by the  
109 unloading caused by the excavation of material following an impact (Michaut  
110 and Pinel, 2018). For a small driving pressure, the magma path is likely to be  
111 controlled by the lithospheric stress state (Watanabe et al., 2002). We model the  
112 stress state below lunar FFCs to quantitatively estimate if and under which condi-  
113 tions it may explain magma flow horizontalization and magma storage at shallow  
114 depth. We thus limit our analysis to FFCs of Class 2, 3 and 4 that show evidence  
115 of a strong deviation of the magma from the vertical. We do not account for FFCs  
116 in the South Pole Aitken Basin (SPA) (Figure 1). GRAIL's gravity data indeed  
117 show regional variations of the crust density of  $\pm 250 \text{ kg m}^{-3}$ , primarily between  
118 the highlands and the SPA (where the crust is denser), which tend to increase the  
119 magma driving pressure and affects its crustal trajectory.

### 120 **3. Stress model**

121 We model the local stress field acting beneath a crater taking into account  
122 two main contributions: i) the unloading forces due to crater formation, and ii) the  
123 loading due to the presence of mare and mascons. The total stress field is obtained  
124 by superposing these two stress fields.



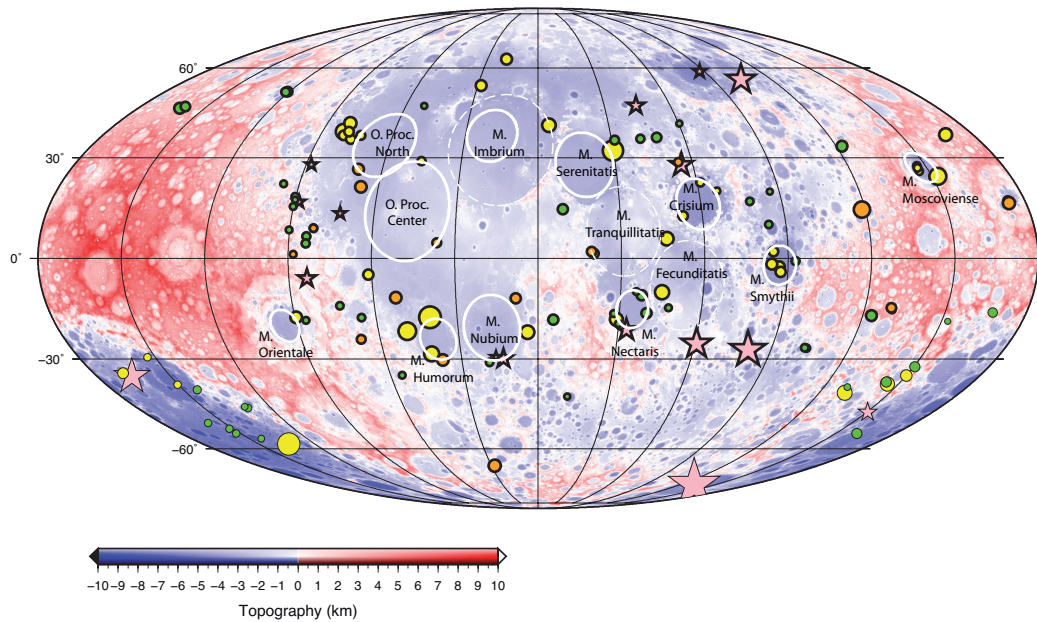


Figure 1: Distribution of lunar floor fractured craters that show an uplifted convex floor (Class 2, orange circles, Class 4 green circles) or an uplifted flat floor (Class 3, yellow circles) or, in pink stars, that show a flooded interior (Class 1) or mare material on their floor (Class 6). The crater distribution is superimposed on the surface topography with Highlands appearing in red. Symbol size is proportional to crater size. Bold symbols represents craters that are used in the following analysis while thin symbols, located in the South Polar Aitken basin area, are not considered. Mare positions and extensions as listed in Table 2 are also represented.

125 *3.1. Local stress due to crater unloading*

126 To calculate the local stress induced by a crater unloading, we follow Michaut  
127 and Pinel (2018) and treat the crater depression as a negative load on an elastic  
128 medium. A bowl-shape axisymmetric crater of radius  $R$  results in a stress  $\sigma_{zz}^C(z =$   
129  $0)$ :

$$\sigma_{zz}^C(0 \leq r \leq R, z = 0) = -\rho_c g h_0 \left(1 - \frac{r^2}{R^2}\right) \quad (1)$$

130 where tensile stresses are assumed negative,  $\rho_c$  is crustal density,  $g$  gravity,  $r$   
131 radial coordinate,  $z$  is the depth defined with respect to the ground plane, and  $h_0$   
132 the maximum crater depth, expressed as a function of the crater radius following  
133 Thorey et al. (2015):

$$h_0 = 540 \times (2 \times R \times 10^{-3})^{0.44} \quad (2)$$

134 where  $h_0$  and  $R$  are in meters. We neglect the loading forces associated with  
135 the crater rim and central peak topography. The stress field caused by the unload-  
136 ing is determined from the equations for linear elasticity which we solved using  
137 the Finite Element Method (COMSOL software) for an elastic plate of finite thick-  
138 ness  $T_e$ , and for a semi-infinite medium. The analytical solutions of Pinel et al.  
139 (2017) for the displacement induced by the loading of a thick elastic plate on an  
140 inviscid medium have been used to validate our numerical solutions. Results for  
141 large plate thicknesses compared to crater radius tend to the analytical results for  
142 a semi-infinite medium (Sneddon, 1951). The vertical profile of radial, tangential  
143 and vertical stresses in a semi-infinite elastic medium of Poisson's ratio  $\nu$  at the  
144 axis ( $r = 0$ ) of a bowl-shape load of radius  $R$  are respectively given by Michaut  
145 and Pinel (2018):

$$\begin{aligned}
\sigma_{rr}^{C,\infty}(z) &= -\frac{\rho_c g h_0}{R^2} \left( (1 + 2\nu) \frac{R^2}{2} - 2(1 + \nu) z \sqrt{R^2 + z^2} + (3 + 2\nu) z^2 - \frac{z^3}{\sqrt{R^2 + z^2}} \right) \\
\sigma_{\theta\theta}^{C,\infty}(z) &= \sigma_{rr}^{C,\infty}(z) \\
\sigma_{zz}^{C,\infty}(z) &= -\frac{2\rho_c g h_0}{R^2} \left( \frac{R^2}{2} + \frac{z^3}{\sqrt{R^2 + z^2}} - z^2 \right)
\end{aligned}$$

146 where the geometry is axisymmetric with respect to  $z$ , the vertical coordinate  
147 oriented downwards, and equal to zero at the surface. The exponent  $\infty$  denotes  
148 a semi-infinite medium. At depth, the horizontal and vertical stresses scale with  
149 the surface unloading and decrease over a characteristic depth similar to the crater  
150 radius.

151 At the unloading axis, the horizontal stress is such that  $\sigma_{rr}^C(0, z) > \sigma_{zz}^C(0, z)$ .  
152 Therefore the direction of maximum compressive stress  $\sigma_1$  is horizontal (Figure 2  
153 top) while the minimum compressive stress  $\sigma_3$  is vertical, and a magma intrusion  
154 with small overpressure would tend to form deep horizontal sills below craters.  
155 Floor appearance at FFCs of Class 2, 3 and 4 are indeed consistent with horizontal  
156 crater-centred intrusions (Thorey and Michaut, 2014) but at shallow depths ( $< 10$   
157 km), allowing for roof uplift.

### 158 3.2. Local stress due to unloading in a regional extension

159 If a pre-existing regional stress field is present, due to the presence of mare  
160 and mascons, the direction of maximum compressive stress below a crater would  
161 result from the superposition of both the mare or mascon loading and the crater  
162 unloading. Since the mare loads are much larger in radius than the elastic litho-  
163 sphere thickness, the vertical stresses can be neglected. If the regional horizontal  
164 state of stress is compressive, the maximum compressive stress along the  $z$  axis

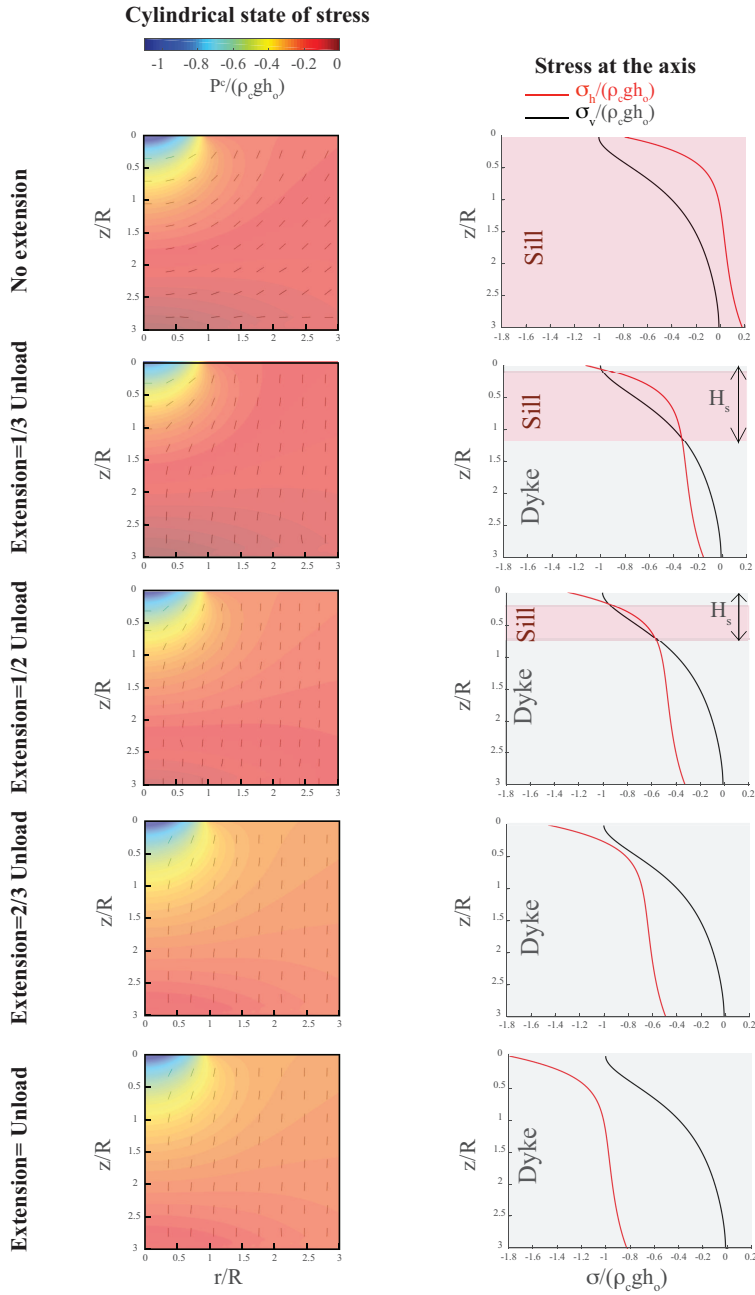


Figure 2: Left, three-dimensional axisymmetric pressure field (colour scale) and direction of maximum compressive stress  $\sigma_1$  (dashes) below a bowl-shape surface unloading of radius  $R$  (equation 1) in an elastic plate of thickness  $T_e = 3R$  and for different pre-existing regional stress fields (top to bottom). The pre-existing regional radial stress and tangential stress  $\sigma_{\theta\theta}^{ext} = \sigma_{rr}^{ext} = \sigma_H^{ext}$  are homogenous vertically and increase progressively in tension from top, where it is  $\sigma_H^{ext} = 0$ , to bottom, where the horizontal stress in tension is equal to the unload  $\sigma_H^{ext} = -\rho_c g h_0$ . Right, for each pre-existing regional stress field, the vertical profiles of radial (red) and vertical (black) stress are shown at the axis of the depression. The depth  $H_s$  where the maximum compressive stress becomes horizontal is indicated. Results are for  $\nu = 0.2$ ,  $E = 10 \text{ GPa}$ ,  $g = 1.62 \text{ ms}^{-2}$ ,

165 remains horizontal. However, if the regional stress displays uniform horizontal  
 166 extension, depending on the magnitude of  $\sigma_H^{ext} < 0$ , the maximum compressive  
 167 stress may become vertical below a given depth  $H_s$  for which  $\sigma_H(z) < \sigma_{zz}(z)$   
 168 (Figure 2). This would allow for vertical dikes deeper than  $H_s$ , and horizontal  
 169 sills above  $H_s$ . The transition depth  $H_s$  progressively becomes shallower as ex-  
 170 tension increases (Figure 2). Actually, mare loading generates a combination of  
 171 both membrane stresses, homogenous with depth, and bending stresses, linearly  
 172 varying with depth, whose effects are considered next.

173 When the regional tensional stress is such that  $\sigma_H^{ext} \leq -0.6\rho_c g h_0$ , the maxi-  
 174 mum compressive stress is vertical over the whole plate thickness (Figure 2, bot-  
 175 tom). This threshold can be derived from the analytical solution for the semi-  
 176 infinite medium and is independent of Young's modulus value. In such a case,  
 177 crater unloading would provide the driving overpressure for magma to rise through  
 178 the crust but no horizontal deflection would occur.

179 Hence, below a relatively small crater or in a thick low density crust (as for  
 180 FFCs of classes 2, 3 and 4), the magma can ascend up to shallow depths ( $< 10$   
 181 km) and then turn into a horizontal sill if the tensional stress field amplitude is  $\lesssim$   
 182  $0.6\rho_c g h_0$ , so that the crater stress field, which favours horizontal sills, dominates  
 183 over extension at shallow depth.

### 184 3.3. *Pre-existing regional stress field due to mascon and mare load*

185 FFCs are distributed around the main maria suggesting either that lunar mag-  
 186 mas were produced only in this specific area due to its enrichment in radioele-  
 187 ments (Laneuville et al., 2013) or that the geological context associated to impact  
 188 basins favoured magma ascent at their periphery because their formation has gen-  
 189 erated pre-existing fractures and magma pathways (Schultz, 1976a) or because

Table 2: Mare characteristics, after <https://planetarynames.wr.usgs.gov>, International Astronomical Union (IAU), Working Group for Planetary System Nomenclature for most mare or Neumann et al. (2015). For simplicity, Oceanus Procellarum (OP) is here modelled as two large circular loads named OP Centre and OP North.

Mare	North Latitude	East Longitude	Radius (km)
Oceanus Procellarum Centre	14	312	450
Oceanus Procellarum North	34	298	300
Mare Serenitatis	28	18	300
Mare Crisium	16	59	230
Mare Smythii	-2	87	177
Mare Moscoviense	27	148	138
Mare Orientale	-20	266	147
Mare Humorum	-24	322	190
Mare Nubium	-21	343	300
Mare Nectaris	-15	35	170
Mascon Imbrium	37	341	250
Mare Imbrium	33	345	530
Mare Tranquillitatis	8	31	400
Mare Fecunditatis	-8	54	400

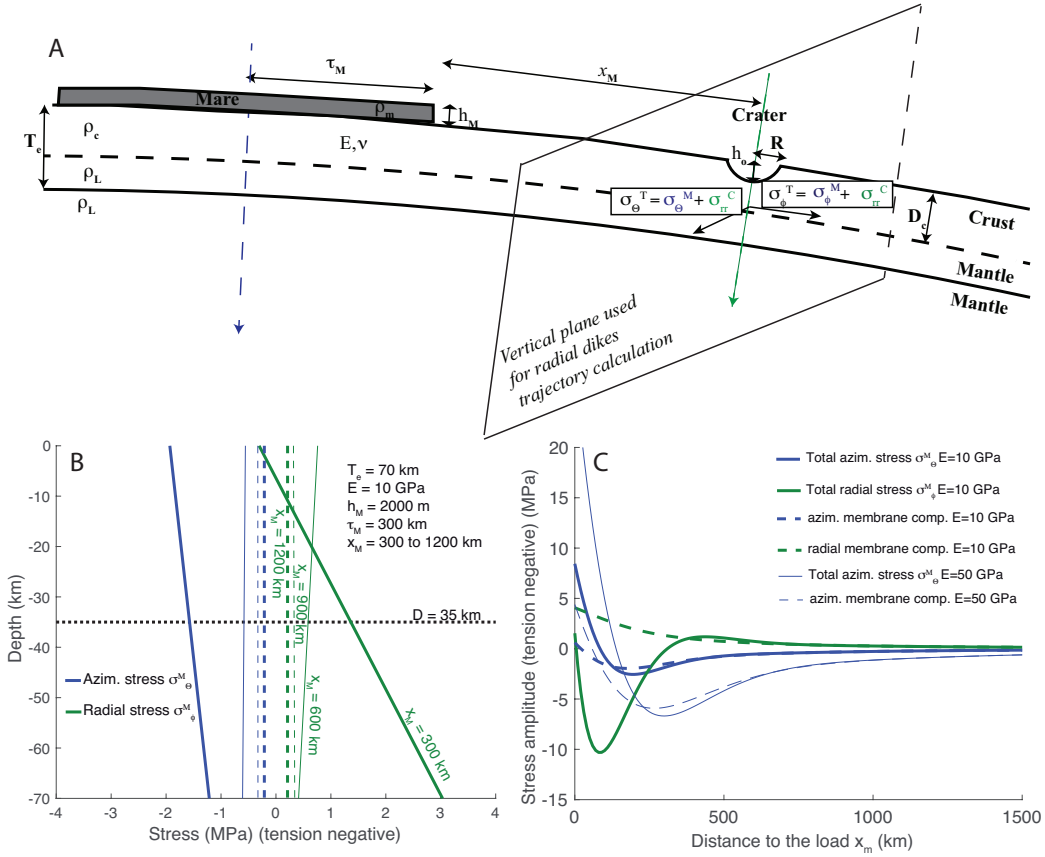


Figure 3: A: Stress field components ( $\sigma_{\theta}^T$  and  $\sigma_{\phi}^T$ ) in the elastic lithosphere at the axis of a bowl-shaped crater of radius  $R$  and depth  $h_o$  at a distance  $x_M$  from a load of radius  $\tau_M$ , thickness  $h_M$  and density  $\rho_M$ . This stress field is the summation of the crater unloading contribution  $\sigma_{rr}^C = \sigma_{\theta\theta}^C$  and the mare/mascon loading one  $\sigma_{\theta\theta}^M$  and  $\sigma_{\phi\phi}^M$ . Note that the elastic lithosphere contains a lithological boundary at depth  $D_c$  separating the crust and the mantle. Bottom: Horizontal azimuthal (blue)  $\sigma_{\theta}^M$  and radial (green)  $\sigma_{\phi}^M$  stresses (with respect to load center) caused by a circular load as a function of depth for four different distances from the load  $x_M = 300, 600, 900, 1200$  km indicated on the graphs (same lines but two colours for azimuthal and radial stress, B) or at the surface as a function of distance from the load  $x_M$  (C). We use  $T_e = 70$  km,  $E = 10 - 50$  GPa,  $\nu = 0.2$ ,  $h_M = 2000$  m,  $\rho_M = 3100$  kg m $^{-3}$ ,  $\rho_L = 3400$  kg m $^{-3}$ ,  $\rho_c = 2500$  kg m $^{-3}$  and  $\tau_M = 300$  km. On C, azimuthal stresses are also indicated in the case  $E = 50$  GPa; membrane stress components (homogenous with depth) dominate at distances  $\geq 3\lambda$  from the load.

190 they induce tensional stresses by lithosphere loading. McGovern and Litherland  
 191 (2011) suggested indeed that lava filling of large circular basins on the Moon cre-  
 192 ated a specific regional stress state that promoted magma ascent through vertical  
 193 dikes in annular regions at the periphery of these large circular loads. We ex-  
 194 amine the possibility, following McGovern and Litherland (2011) and Thomas  
 195 et al. (2015), that lithosphere loading by mascons and mare weight has induced a  
 196 pre-existing regional extensive stress on their periphery, favouring magma ascent  
 197 through vertical dikes to shallower depths.

198 We model the regional lithostatic stress state  $\sigma^M$  induced by a mare load using  
 199 the analytical solution for azimuthal and radial stress  $\sigma_{\Theta}^M$  and  $\sigma_{\Phi}^M$  around a circular  
 200 load (Brotchie, 1971; Solomon and Head, 1979; Freed et al., 2001). We account  
 201 for bending and membrane stresses in an elastic spherical shell of thickness  $T_e$   
 202 (see Appendix).

203 The amplitude of regional horizontal stresses are sensitive to the load prop-  
 204 erties: assuming a circular mare load, they increase with mare radius  $\tau_M$  (Figure  
 205 3) and are directly proportional to its surface mass or load thickness  $h_M$ , for a  
 206 given load density. Bending stresses vary linearly with depth and dominates over  
 207 membrane stresses closer to the mare over a width that is equal to a few times the  
 208 flexural wavelength of the elastic lithosphere  $\Lambda = \left( \frac{ET_e^3}{(1-\nu^2)\rho_L g} \right)^{1/4}$ , where  $E$  and  $\nu$   
 209 are the lithosphere Young's modulus and Poisson's ratio,  $\rho_L$  is lithosphere density.  
 210 For  $T_e = 70$  km,  $E = 10$  GPa,  $\Lambda \approx 85$  km and bending stresses dominates over  
 211 a width of  $\sim 0 - 250$  km (Figure 3). Very close to the load, bending causes ten-  
 212 sional radial stresses close to the surface and tensional azimuthal stresses at depth  
 213 ( $\sim 0 - 250$  km from the load on Figure 3C) and inversely a little farther away  
 214 ( $\sim 250 - 350$  km from the load on Figure 3C). Membrane stress components are



215 constant with depth in the lithosphere and are dominant in the far field ( $> 200$ -  
216 300 km from the load, Figure 3C). Radial membrane stresses are in compression;  
217 on the contrary, azimuthal membrane stresses are in tension (Figure 3, B and C),  
218 favouring radial fracture opening. The azimuthal membrane stress decreases from  
219 a small positive value (compression) at the load border to a minimum negative  
220 value (maximum in tension) at  $\sim 200$  km from the load for the example of Figure  
221 3C; it then slowly increases back towards zero at larger distances. The azimuthal  
222 membrane stress caused by mare loading could have favoured radial dyke opening  
223 over distances larger than a thousand kilometers from the load. However, it was  
224 not the cause of magma ascent below FFCs, since, in that case, the magma would  
225 have reached the surface everywhere and not formed crater-centred intrusions.

#### 226 3.4. Total stress field

227 To compute the total stress field acting beneath the crater, we superpose the  
228 local stress field induced by the crater, estimated using COMSOL numerical cal-  
229 culations (Figure 2 top), and the pre-existing regional stress field caused by a  
230 circular mare load on an elastic spherical lithosphere of finite thickness  $T_e$  (Figure  
231 3). To examine a wide range of parameters, we estimate the stress field at the  
232 crater axis and look for solutions favourable to shallow crater-centred sill intru-  
233 sions. In terms of stress state, solutions should allow for the formation of a dike  
234 (radial or concentric relative to mare geometry) at the base of the crust  $D_c$ , hence  
235 a horizontal least compressive stress at  $z = D_c$ , as well as the formation of a sill  
236 at shallower depths in the crust, i.e. a vertical least compressive stress at a depth  
237  $z$  such that  $0 < z < D_c$  (Figure 4A). The magma is assumed buoyant in the litho-  
238 spheric mantle where its path is not affected by the stress field. Solutions depend  
239 on crater radius  $R$ , lithosphere thickness  $T_e$ , distance to load  $x_M$ , load thickness

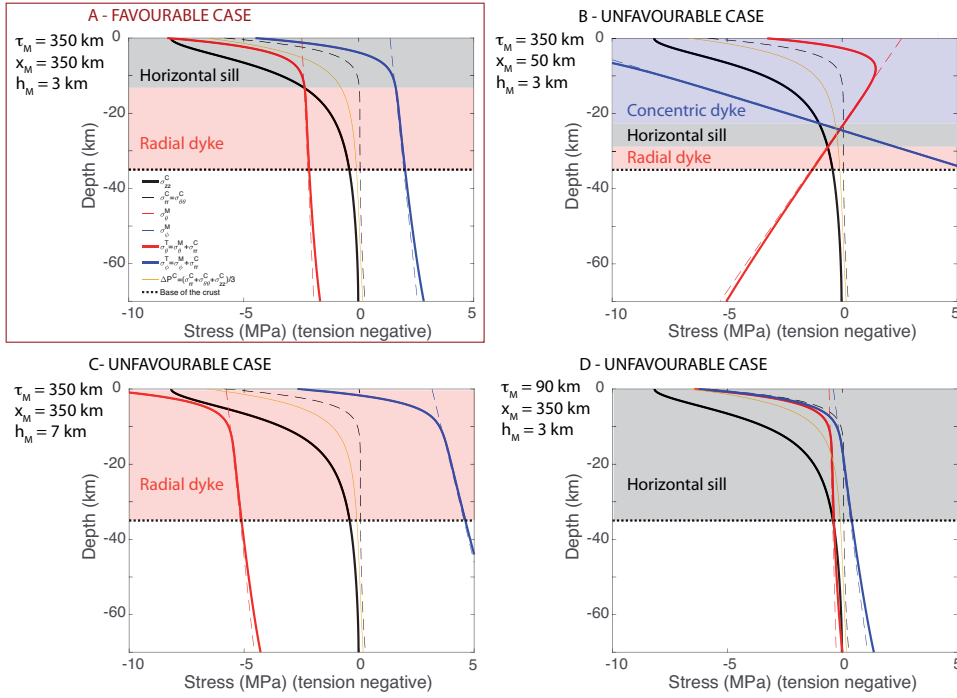


Figure 4: Different cases of stress field at the axis of an unloading as a function of depth for  $T_e = 70$  km,  $E = 10$  GPa,  $R = 10$  km, and for different indicated values of load radius  $\tau_M$ , distance to mare border  $x_M$  and load thickness  $h_M$ . Black solid line: vertical stress, only due to crater unloading. Red solid line: total azimuthal stress with respect to mare centre, sum of the azimuthal stress due to mare load (red dashed line) and horizontal stress due to crater unloading (black dashed line). Blue solid line: total radial stress with respect to mare centre, sum of the radial stress due to mare load (blue dashed line) and horizontal stress due to crater unloading (black dashed line). Yellow solid line: pressure perturbation due to unloading. An arbitrary crustal thickness of 35 km is illustrated by a horizontal dotted line. Only the stress state in A allows for vertical magma ascent followed by flow horizontalization at a shallow depth. In B, vertical dike is favoured at the base of the crust followed by a sill at shallower depths but bending stresses are so important close to the load ( $x_M = 50$  km) that the total tectonic stress gradient opposes to magma ascent and cannot be compensated by the driving pressure due to unloading. In C, extension induced by mare loading dominates preventing sill formation whereas in D it is too small to enable dike formation at the crust base.

240  $h_M$ , load radius  $\tau_M$  as well as Young's modulus value. Depending on the values  
 241 of these parameters we obtain four different stress regimes (Figure 4, panels A to  
 242 D), which would cause different propagations of magmatic intrusions nucleated  
 243 at the crust base below the crater.

244 If the amplitude of the azimuthal membrane stress is larger than  $0.6\rho_cgh_0$ , it  
 245 dominates over the crater stress field, which would cause magma ascent in radial  
 246 dikes and eruption within the crater floor (Figures 4C and 2 bottom). Such a  
 247 scenario would not explain observed floor uplift at FFCs of Classes 2, 3 and 4. On  
 248 the contrary, if the membrane stress is too small, such that a vertical dyke is not  
 249 allowed at the base of the crust, the magma would form deep horizontal intrusions,  
 250 which would not explain floor deformations either (Figure 4D).

251 Close to the load ( $x_M \lesssim 200$  km), bending stress dominates over membrane  
 252 stress and might be such that a vertical dike is favoured at the crust base followed  
 253 by a horizontal sill within the crust (Figure 4B). However, in that case, the vertical  
 254 stress gradient  $d\sigma^H/dz$  is positive and opposes to magma ascent (the azimuthal  
 255 pre-existing stress largely increases with height, red curve on Figure 4B). The  
 256 overpressure  $P^C$  provided by crater unloading at  $z = D_c$  allows for magma ascent  
 257 to shallow depths ( $< 10$  km) only if  $P^C$  counteracts the positive vertical gradient  
 258 in total stress up to 10 km depth, or up to the depth at which the vertical gradient  
 259 in total stress reverses ( $z^{\theta\phi} > 10$  km) favouring magma ascent:

$$\begin{aligned}
 |P^C(z = D_c)| &\geq \sigma_{\Phi}^T(z = \max(z^{10}, z^{\theta\phi})) - \sigma_{\Phi}^T(z = D_c) & \text{if } \sigma_3(D_c) = \sigma_{\Phi}(D_c) \\
 |P^C(z = D_c)| &\geq \sigma_{\Theta}^T(z = \max(z^{10}, z^{\theta\phi})) - \sigma_{\Theta}^T(z = D_c) & \text{if } \sigma_3(D_c) = \sigma_{\Theta}(D_c)
 \end{aligned}
 \tag{3}$$

260 where  $\Theta$  and  $\Phi$  are respectively the azimuthal and radial coordinates relative to  
 261 the load centre,  $\sigma_3$  the least compressive stress,  $\sigma_{\Phi}^T = \sigma_{\Phi}^M + \sigma_{rr}^C$  and  $\sigma_{\Theta}^T = \sigma_{\Theta}^M + \sigma_{rr}^C$ ,  
 262 where  $\sigma^M$  is the stress generated by mascon/mare loading on the lithosphere and  
 263  $\sigma^C$  the stress perturbation due to crater unload.

## 264 4. Results

### 265 4.1. Conditions favourable to shallow crater-centred magmatism

266 For a given load radius, lithosphere elastic thickness and Young's modulus  
 267 value, we look for solutions in terms of crater radius  $R$ , load thickness  $h_M$  and  
 268 distance to load  $x_M$  that would favour: i) a vertical dike at the crust base, ii) a  
 269 dike to sill transition within the crust as well as iii) an effective overpressure that  
 270 would drive magma up to shallow depths  $\leq 10$  km (Equations 3).

271 For distances  $\gtrsim 200$ -300 km, membrane stresses are dominant. Results show  
 272 that as the mare thickness increases, the condition for having FFCs are met at  
 273 increasing distances from the mare, and within a wider range. This result stands  
 274 given a fixed load radius, crater radius and lithosphere thickness (Figure 5). When  
 275 the load is too thick (for instance for  $h_M > 4000$  m for  $R = 10$  km,  $T_e = 70$   
 276 km,  $E = 10$  GPa and  $\tau_M = 350$  km), azimuthal membrane stresses close to the  
 277 load (at  $x_M \leq 200 - 300$  km) become larger than  $0.6\rho_c g h_0$  and dominate over the  
 278 local stress due to crater unloading (Figures 4C and 5). Far away from the load,  
 279 membrane stresses become too small and vertical dikes cannot form at the base of  
 280 the crust leading to deep sill intrusions (Figures 4 and 5).

281 At distances  $\lesssim 200$  km to the mare, bending stresses are dominant and tec-  
 282 tonic stresses tend to oppose to magma ascent. For the magma to ascend to shal-  
 283 low depths, the overpressure due to crater unloading has to compensate for the

284 positive tectonic stress gradient. This is only possible if load thicknesses are suffi-  
285 ciently small. As the crater radius, and hence the induced overpressure, increases,  
286 the maximum possible load thicknesses slowly increases for distances  $\lesssim 200$  km  
287 (Figure 5).

288 Our results show that, for a given mare thickness and radius, the smaller the  
289 crater radius, the easier it is for membrane stress to compensate for the vertical  
290 extension at depth caused by crater unloading. Therefore, shallow magmatism  
291 can occur below the crater at increasing distance from the mare (Figure 5) for a  
292 given load thickness and radius. As a result, shallow horizontal intrusions beneath  
293 small craters ( $R < 10$  km) are likely to occur over a larger range of distances from  
294 the mare. Instead, shallow intrusions below large craters ( $R > 20$  km) should only  
295 occur close to the mare (Figure 5).

296 Indeed Schultz (1976a) noted that small FFCs are found deep into the High-  
297 lands. To quantify this observation, we calculate the distance of each FFC to its  
298 closest and second closest mare by considering the main maria and assuming cir-  
299 cular mare loads with a given centre and radius (see Table 2, Figure 1). Such a  
300 geometry is certainly simplified, in particular for Oceanus Procellarum (OP) that  
301 represents a broad positive gravity anomaly. But the 3D loading geometry is more  
302 difficult to represent. Mare Orientale is also associated to a more complex gravity  
303 signature, with a central gravity high encircled by a low gravity annulus itself sur-  
304 rounded by a broader gravity high region, which might be important to consider  
305 in the interpretation of individual FFC. Given that membrane stress plays over a  
306 large distance from the mare, it may also happen that the mare which is respon-  
307 sible for magma ascent is not the closest to the crater. However, our simplified  
308 model allows to obtain first-order quantitative results, and test our assumptions.

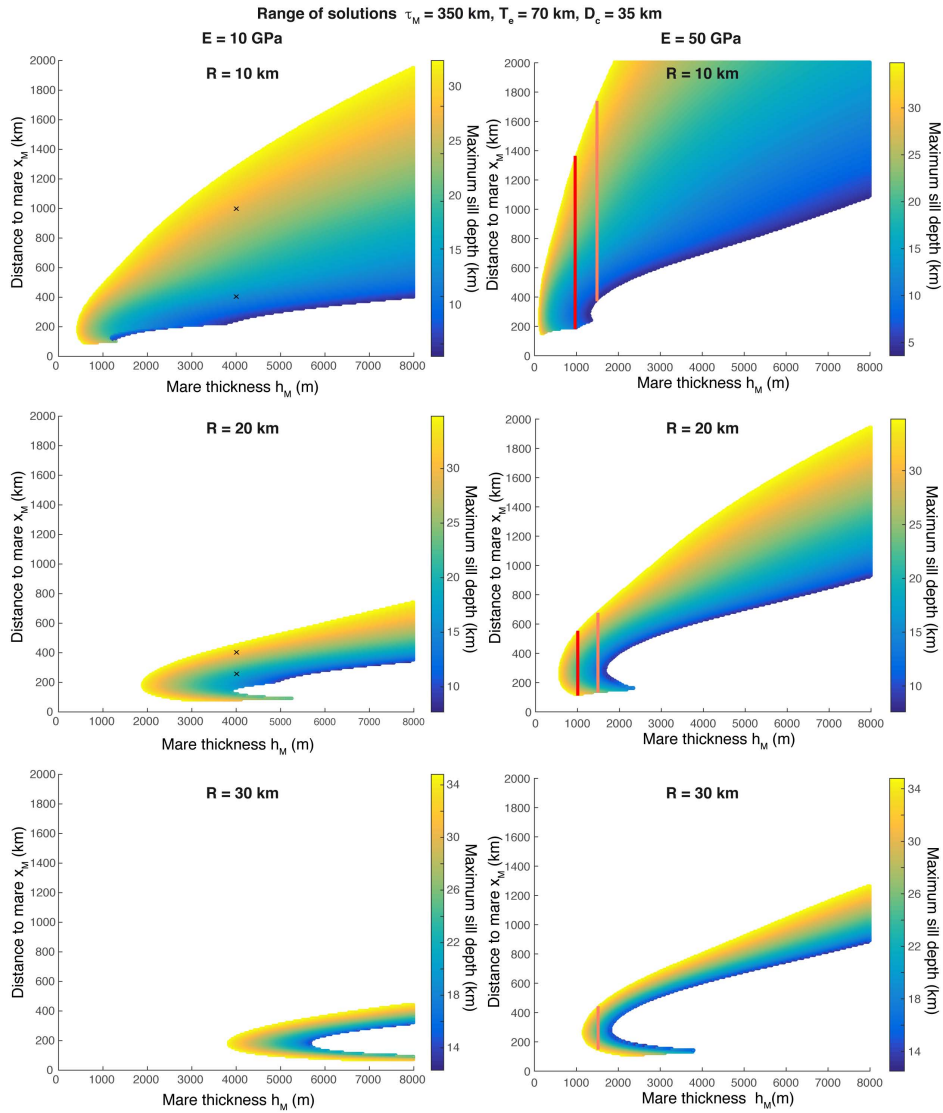


Figure 5: Range of possible solutions (coloured area) in terms of distance to mare load  $x_M$  and mare thickness  $h_M$  for a shallow magma intrusion to occur below a crater for different crater radius of 10 (top), 20 (middle) and 30 km (bottom), for a load radius  $\tau_M = 350$  km, an elastic lithosphere thickness  $T_e = 70$  km and a crustal thickness  $D_c = 35$  km and using  $E = 10$  GPa (left), or  $E = 50$  GPa (right).  $\nu = 0.2$ ,  $g = 1.62 \text{ ms}^{-2}$  and  $\rho_L = 3400 \text{ kg.m}^{-3}$ ,  $\rho_c = 2500 \text{ kg m}^{-3}$ ,  $\rho_M = 3100 \text{ kg m}^{-3}$ . The maximum sill depth is the maximum depth at which the minimum compressive stress is vertical at the crater axis, thus assuming that the magma trajectory follows  $\sigma_1$ . The range of distances indicated in red for  $h_M = 1000$  m and orange for  $h_M = 1500$  m for  $E = 50$  GPa correspond to the ones indicated on Figure 6A. Crosses on the left ( $E = 10$  GPa) indicate the distance, mare load and crater radius used to calculate the stress state necessary to 2D simulations of Figure 7.

309 Results show that the range of observed distance to the mare indeed decreases  
310 as the crater radius increases: while small craters of radius  $\sim 10$  km are found up  
311 to 1200 km away from a mare side, larger craters of  $\sim 30$  km are found within a  
312 few hundreds of km from the maria (Figure 6A). A linear regression with negative  
313 slope fits FFCs radii as a function of their distance from the closest maria (Figure  
314 6). Distances are considered with an error  $\Delta x_M = \pm 100$  km. Considering or not  
315 Mare Tranquillitatis, Fecunditatis and the whole extent of Mare Imbrium do not  
316 significantly change the results (Figure 1), suggesting that these maria were not  
317 important actors for the occurrence of shallow magmatism below craters.

318 Figure 5 also shows the depth at which the minimum compressive stress turn  
319 from horizontal to vertical. In case the magma trajectory exactly follows  $\sigma_1$ , this  
320 would indicate a dyke to sill transition. These depths might be quite large, in  
321 particular far from the load. However, depending on the trade-off between the  
322 magnitude of the regional stress and the magma driving force, intrusions may  
323 display a delayed deflexion towards the direction of  $\sigma_1$  (Watanabe et al., 2002;  
324 Menand et al., 2010), which would validate the whole possible range of solutions  
325 of Figure 5. This assumption is tested in the following section.

#### 326 4.2. *Results from dike propagation model*

327 In our analysis, we assumed that buoyancy and overpressure were small enough  
328 for the magma to exactly follow the direction of the maximum compressive stress.  
329 This approach has allowed us to look for solutions over a wide range of param-  
330 eters in terms of load characteristics and distance from the mare. In order to  
331 validate these results, we make use of a 2D boundary element (BE) model to sim-  
332 ulate the trajectory of a magmatic dike propagating within the crust (Maccaferri  
333 et al., 2011). Results from the BE model for dike propagation allow us to quantify

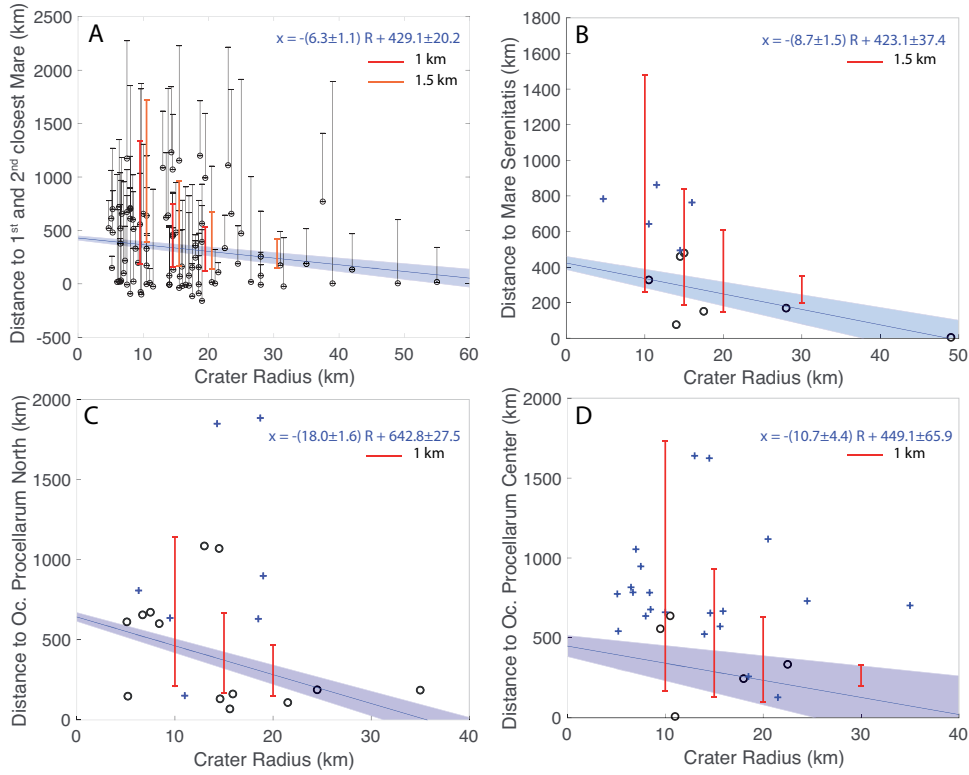


Figure 6: A: Distance from first (circle) and second (dash) closest mare for each floor fractured craters of Class 2, 3 and 4 (showing an elevated floor) as a function of crater radius. We assume circular maria of given radius and centre (Table 2). Thick lines indicate predicted distance range of FFCs of given radius from a load of radius  $\tau_M = 350$  km and thickness 1 (red) or 1.5 km (orange). The linear regression line and equation fitting data corresponding to the closest mare are shown in blue, with the coloured area corresponding to the regression error for an error on the distance of  $\pm 100$  km. B, C, D: Distance of closest FFCs from different maria as a function of FFCs radius. Black circles: FFCs for which the mare considered is the closest one, blue crosses: FFCs for which the mare considered is the second closest one. Predicted distance ranges are indicated as errorbars in red and the regression line (regression error is obtained considering an error on the distance of  $\pm 50$  km) and its equation fitting data corresponding to FFCs for which the Mare is the closest one (circles) are indicated in blue and coloured area. B: Mare Serenitatis, modelled as a cylindrical load of radius 300 km. Predicted distance ranges are for a load radius of 300 km and thickness 1.5 km. C: Northern part of OP, modelled as a cylindrical load of radius 300 km. Predicted distance ranges are for a load of radius 300 km and thickness 1 km. D: Central part of OP modelled as a load of radius 450 km. Predicted distance ranges are for a load radius of 450 km and thickness 1 km. Calculations use  $E = 50$  GPa,  $T_e = 70$  km,  $D_c = 35$  km,  $\nu = 0.2$ ,  $g = 1.62$  ms $^{-2}$  and  $\rho_L = 3400$  kg.m $^{-3}$



334 the height required for a vertical dike to turn into a sill. Using this approach we  
335 are able to drop the assumption that dikes instantaneously adjust to the direction  
336 favoured by the principal stresses. Dike trajectories are computed by maximising  
337 the strain plus gravitational energy released by the dike on incremental propa-  
338 gations along different test directions (Maccaferri et al., 2011). This calculation  
339 allows to track the actual dike trajectory which does not depend only on the local  
340 stress field but also on magma driving force and to check under which conditions  
341 a vertical dike starting from the base of the crust may get stacked as a horizontal  
342 sill at shallow depth beneath an impact crater. The BE model for dike propagation  
343 is fed with a radial cross section of the stress field acting within the crust beneath  
344 an impact crater, as computed with COMSOL (Figure 5). The horizontal stress  
345 component is given by  $\sigma_{\Theta}^T$  and the vertical component is only induced by crater  
346 unloading.

347 We consider several specific favourable cases where the membrane stress com-  
348 ponent induced by mare loading dominates, producing azimuthal tensile stress and  
349 favouring radial dikes turning into sills. Vertical dykes start 10 km below the crust,  
350 at 45 km depth, with a given volume and internal pressure, at different distances  
351 ( $x_C = 0.5, 1.0, 2.0, 30$  km) from the crater axis (Figure 7). Magma is assumed to  
352 be buoyant in the mantle but not in the crust (which is 35 km thick).

353 Although magma loses pressure in the crust, where it is negatively buoyant,  
354 its driving pressure allows for further ascent and the intrusion does not stop at the  
355 crust base. Simulations show that for larger magma volume and driving force,  
356 the dyke is less deviated from its vertical path and ascends farther up in the crust.  
357 Therefore, the transition from a dyke to a sill occurs at shallower depths than  $H_s$ .  
358 If the driving force is sufficiently large (Figure 7), the magma might even reach

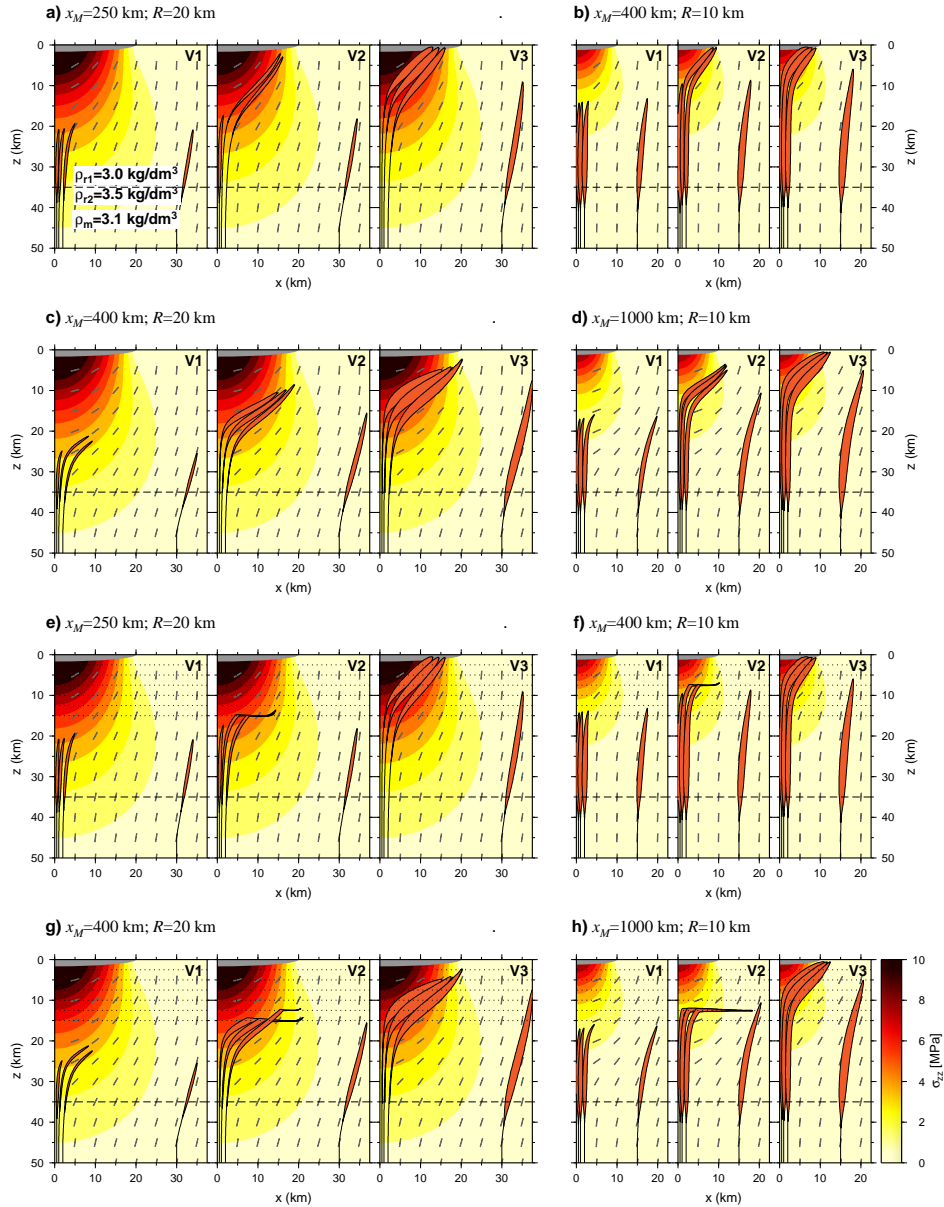


Figure 7: Panels 'a' to 'd') Simulated dike trajectories (red elongated sheet intrusions) for different stress scenarios, and different magma volumes (that is the dike cross section, in the 2D model:  $V_3 = 0.25 \text{ km}^2$ ,  $V_2 = 0.5 \times V_3$ ,  $V_1 = 0.5 \times V_2$ ). Panels 'e' to 'h') same simulations but including six weak layers (dotted lines) at regular depth intervals between 2.5 and 15 km depth. In all simulations dikes start at 45 km depth.  $\rho_{r1}$ ,  $\rho_M$ , and  $\rho_{r2}$  are the densities of crustal rocks, magma, and mantle rocks, respectively. The dashed horizontal line is the crust-mantle boundary. Grey dashes indicate the direction of  $\sigma_1$ . The intensity of the vertical component of loading and unloading stresses is plotted as colour contour. In panels a) and e) we consider the stress induced by a crater of radius  $R = 20 \text{ km}$  at a distance  $x_M = 250 \text{ km}$  from the load. b) and f)  $R = 20 \text{ km}$ ,  $x_M = 400 \text{ km}$ . c) and g)  $R = 10 \text{ km}$ ,  $x_M = 400 \text{ km}$ . d) and h)  $R = 10 \text{ km}$ ,  $x_M = 1000 \text{ km}$ . We used  $T_e = 70 \text{ km}$ ,  $\tau_M = 350 \text{ km}$ ,  $h_M = 4 \text{ km}$ ,  $E = 10 \text{ GPa}$  and  $\nu = 0.2$ .

359 the surface, as also shown by Michaut and Pinel (2018) for large craters on a  
360 thin crust. Far from the crater axis, the local crater stress field is negligible and  
361 the vertical dyke is not deviated. Dykes starting far away from a crater would also  
362 stagnate at deeper depths than the ones starting below the crater axis. This is due to  
363 the larger decrease in pressure as the intrusion ascends below the crater unloading  
364 (Michaut and Pinel, 2018).

365 Our calculations show that the depth of the dyke-to-sill transition is controlled  
366 by the magnitude of the deviatoric stress compared to magma driving force. The  
367 deviatoric stress below craters becomes significant at a depth similar to the crater  
368 radius (Figure 2). Hence, below small craters on a thick crust, if the magma intru-  
369 sion starts as a vertical dyke, it can ascend up to shallow depths equivalent to the  
370 crater radius before being deviated (Figure 7). This confirms that the entire range  
371 of distances from the mare for which we expect FFCs, according to the calcula-  
372 tions carried out in the previous section, indeed lead to shallow sill intrusions.

373 Finally, we test the effect of a weak layer interface on magma propagation,  
374 to reproduce the presence of a brecciated or fractured zone, which has been pre-  
375 viously proposed as a possible cause for sill emplacement below impact craters  
376 (Schultz, 1976a; Jozwiak et al., 2012). Our simulations show that such a weak  
377 layer can help the process of horizontalization if the magma driving force is low  
378 enough (for a large volume or large driving pressure, the dyke path always remains  
379 unperturbed by the weak layer, see cases V3, bottom of Figure 7). However, in  
380 order for the dyke to intrude through the weak interface (characterised by null  
381 fracture toughness of the rocks), it is necessary to consider also the local stress  
382 field due to the crater unloading. In fact, dykes starting far from a crater, which  
383 are not directly affected by the crater unloading, does not intrude within the weak

384 interface (Figure 7 panels f and h volume V2).

### 385 4.3. *Effect of Young's modulus, crustal and elastic lithosphere thicknesses*

386 For a given elastic plate thickness  $T_e$ , crater unloading causes compression  
387 at depths  $z \gtrsim 2T_e/3$  which inhibits magma ascent (Figure 3 in Michaut and  
388 Pinel (2018)). The presence of crater-centred intrusions below craters in region of  
389 crustal thickness of  $\sim 30 - 40$  km (Figure 3 in Michaut and Pinel (2018)) thus ar-  
390 gues for a relatively thick lithosphere such that  $T_e \geq 70$  km at the time of magma  
391 ascent below craters. The Young's modulus value has a negligible effect on this  
392 estimate (in particular, Equation 3 does not depend on  $E$ ).

393 As long as  $D_c \lesssim 2T_e/3$ , increasing the crustal thickness would increase the  
394 maximum distance from the mare at which FFCs can form. This is because a  
395 thicker crust would more easily allow for vertical dykes at its base, wherever the  
396 effect of mare loading dominates over crater unloading (Figure 2). The fact that  
397 the crust generally thickens farther from the mare (Wieczorek et al., 2013) thus  
398 favours the development of crater-centred intrusions far from the load.

399 For a given crustal thickness, crater and load radius, a thicker elastic litho-  
400 sphere slightly increases the minimum load thickness necessary to meet the con-  
401 dition for FFCs formation. However, the effect of elastic thickness variations on  
402 possible distance range of FFCs from mare is negligible.

403 Young's modulus values are larger for mantle rocks than for crustal rocks and  
404 decrease with porosity and temperature (Pritchard and Stevenson, 2000) spanning  
405 a range between 10 and 100 GPa, with most studies considering values between  
406 50 and 100 GPa for the lithosphere (Solomon and Head, 1979). The value of  $E$   
407 quite significantly affects our results: the larger  $E$ , the larger the amplitudes of  
408 both bending and membrane stresses due to mare loading are (see Appendix) and

409 the wider the range of possible distances for FFCs from the load for given load  
410 properties (i.e. given  $\tau_M$ ,  $h_M$ ), (Figure 5). In the following, we consider a value  
411 of 50 GPa and  $T_e = 70$  km.

#### 412 4.4. Mare thicknesses

413 The main maria have radii between  $\sim 150$  and 450 km (see Table 2).

414 Within this range of load radius, we can calculate the minimum thickness that  
415 is necessary for magma ascent up to the floor of a crater with radius of 10 km.  
416 For this calculation, we use both the minimum value of the azimuthal membrane  
417 stress, that occurs at a few hundreds of kilometers from the load (Figure 3 B  
418 and C), as well as the mean value over the distance range 100-700 km from the  
419 load, that is the range of distances over which most FFCs of radius  $\sim 10$  km are  
420 observed (Figure 6 A).

421 Results show that, in order to have small uplifted FFCs ( $\sim 10$  km in radius),  
422 at a distance between 100 to 700 km from a load with radius  $> 250$  km, the load  
423 thickness must be smaller than  $\sim 3$  km (Figure 8). Smaller FFCs, of Class 4  
424 in particular, show uplifted floors rather than flooded floors. These craters are  
425 observed around the largest maria, in particular OP, Mare Serenitatis and Mare  
426 Crisium. Given mare radius (see Table 2), we deduce that the mare thickness is on  
427 average less than  $< 1.6$  km in OP in its central part,  $< 2.3$  km in Mare Serenitatis  
428 and  $< 3.4$  km in Mare Crisium.

429 Taken all together, the range of distances between all FFCs with an uplifted  
430 floors and their first and second closest mare is consistent with an average mare  
431 thickness of  $\sim 1$  to 1.5 km, assuming  $\tau_M = 350$  km (Figure 6 A).

432 FFCs may be influenced by a set of stresses imposed by the presence of more  
433 than one mare. Here, we assume that a FFC is under the influence of a given mare

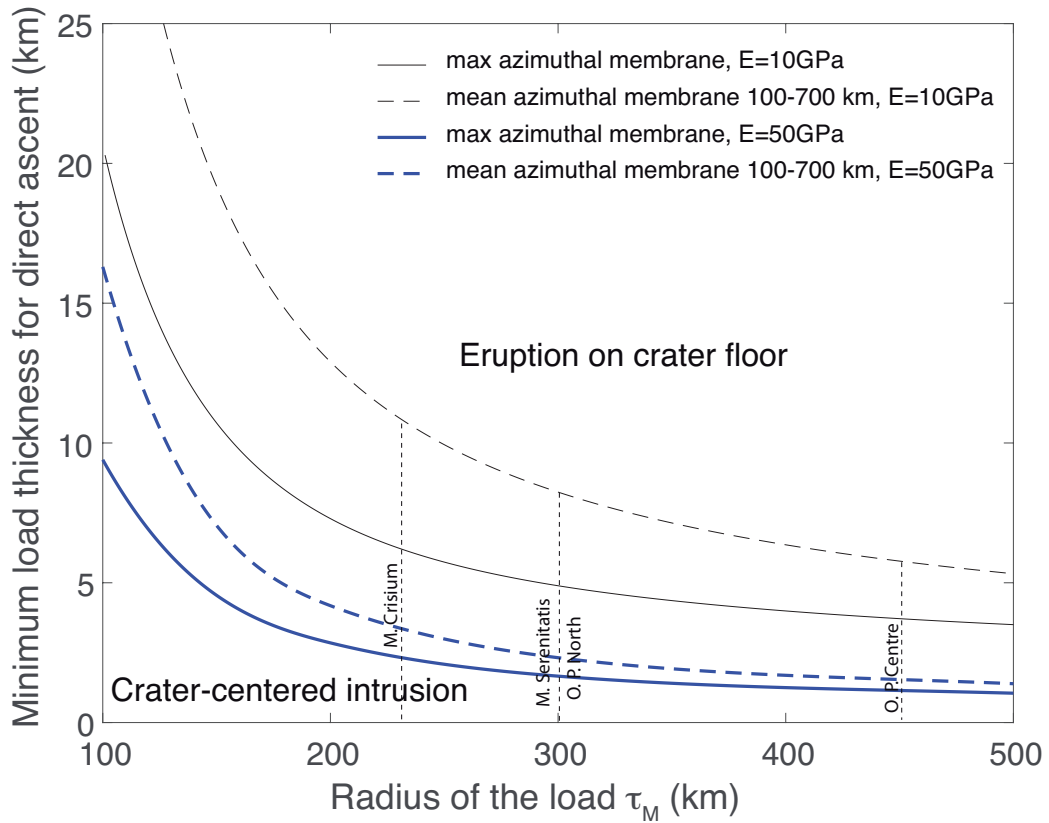


Figure 8: Load thickness  $h_{asc}$  necessary for a direct ascent up to the crater surface for a crater radius of 10 km as function of load radius  $\tau_M$ . Dashed line: thickness calculated using the mean value of the azimuthal membrane stress between 100 and 700 km from the load. Solid line: thickness calculated using the maximum azimuthal membrane stress in tension (i.e. minimum value). Black lines:  $E = 10$  GPa, blue lines,  $E = 50$  GPa.

434 or mascon if this mare is the first or second closest mare to this crater. We observe  
435 that the distance range of FFCs from the mare indeed decreases as the FFC radius  
436 increases (Figure 6). In fact, a linear regression with significant negative slope fits  
437 FFCs radii versus their distance from the closest mare (Figure 6 B, C and D).

438 We found that FFCs distribution around OP are in general consistent with  
439 an average mare thickness of 1 km, in agreement with results from Gong et al.  
440 (2016), based on GRAIL's gravity data, but about twice the value proposed by De  
441 Hon (1979). However, the method used by De Hon (1979) only provides lower  
442 bounds on basalt thicknesses (Gong et al., 2016) and these authors also noted local  
443 lenses in excess of 1000-1500 m which increases the average load thickness.

444 The distribution of FFCs around Mare Serenitatis is consistent with a load  
445 thickness of  $\sim 1.5$  km for a load radius of 300 km. This estimate is in good agree-  
446 ment with Michaut et al. (2016) who found a thickness larger than 1.3 km and  
447 with the estimation of 1.6 km based on the deepest reflector detected by the Lunar  
448 Sounder Experiment (Peeples et al., 1978). East of Mare Serenitatis, Weider et al.  
449 (2010) estimated the thickness of 6 different layers, for a total of about 1500 m,  
450 also consistent with this study.

451 Evans et al. (2016) interpreted circular gravity anomalies lacking topographic  
452 expressions as flooded craters. By estimating their initial rim height, they obtained  
453 thicknesses larger than 1.5 km. Their study also point to local basalt lenses in  
454 excess of 7 km thickness. It appears quite logical that our regional-scale study fits  
455 better with their average value. We notice that our FFCs are generally observed  
456 closer to mare than what is expected from our model (Figures 6 and 8). Such a  
457 discrepancy might be explained by the thickening of the load towards the center  
458 of the mare, which is not considered in our simplified loading model.

## 459 **5. Discussion and conclusion**

460 Recently, Michaut and Pinel (2018) have shown that crater unloading may  
461 have triggered magma ascent within the lunar crust below mare-filled and floor-  
462 fractured craters. For large craters, the unloading was large enough to cause  
463 magma eruption on the crater floor. For small craters, the crater unloading was  
464 not large enough to cause an eruption and resulted in an intrusion. In that case,  
465 the magma driving pressure is small and the magma path is likely sensitive to the  
466 lithospheric stress state. Here we demonstrate that flow horizontalization, leading  
467 to sill-like intrusions below FFCs, was caused by the local stress state below crater  
468 unloading with a vertical least compressive stress favouring the emplacement of  
469 horizontal intrusions. The presence of a highly fractured, brecciated zone caused  
470 by the impact over a depth proportional to the crater size and possibly capped by  
471 impact-generated solidified melt (Schultz, 1976a; Wichman and Schultz, 1995),  
472 that we modelled as a weak layer interface below the crater, could have helped,  
473 but did not cause, the horizontalization of magmatic intrusions.

474 Furthermore, a tensional stress is necessary to explain horizontal spreading at  
475 shallow depth and crater floor uplift at FFCs. Mare and mascons loading on the  
476 lithosphere caused tensional membrane stresses in the azimuthal direction for a  
477 circular mare at their periphery over distances larger than  $\sim 1000$  km favouring  
478 radial dyke opening. We demonstrate that the size and spatial distribution of FFCs  
479 of Class 2, 3 and 4, with the smaller FFCs being the farther away from the mare,  
480 is consistent with magma path below craters being controlled by the lithospheric  
481 state of stress, sum of the local stress due to crater unloading and of a regional  
482 horizontal pre-existing stress caused by mare loading. The distribution of FFCs  
483 around lunar maria is consistent with elastic lithosphere thicknesses  $\geq 70$  km and



484 average mare thicknesses  $< 4$  km, in particular of  $\sim 1$  km in OP using  $E = 50$   
485 GPa.

486 The conditions for which magma may accumulate as horizontal sills at shal-  
487 low depths ( $z \leq 10$  km) below impact craters are likely met below small craters  
488 ( $R \leq 10$  km) surrounding mascons and maria. Indeed, the smaller the crater,  
489 the smaller the required additional pre-existing extension to obtain a vertical  $\sigma_1$   
490 at depth ensuring the ascent of magma to shallower depths. It is thus likely that  
491 the mechanism we propose is also the cause of the formation of lunar concentric  
492 craters (Trang et al., 2016). These craters, that have diameters commonly  $< 15$   
493 km, present anomalously shallow floors and a concentric ridge; they share the  
494 same distribution, surrounding the maria, as floor-fractured craters and probably  
495 result from crater-centred intrusions (Trang et al., 2016; Schultz, 1976b).

496 Here, we assume that the stress field seen by ascending magma is the sum-  
497 mation of the regional mare-controlled stress before impact and of the unloading  
498 stress associated with crater topography. However, material fluidization caused  
499 by impact could reset the stress state in the crater subsurface, losing the memory  
500 of prior regional stress within an hemisphere of radius  $\sim R$ . In that case, once  
501 the crater motion ceases, the elastic medium still has to compensate for the final  
502 crater topography, and unloading forces would still dominate the local stress, pro-  
503 moting shallow sill formation (vertical  $\sigma_3$ ). As the stress will not be discontinuous  
504 after the impact and the regional stress is still present at depths  $> R$ , the resulting  
505 stress at shallower depths should be equivalent to a reduced regional stress field in  
506 addition to the local crater stress field. In that case, our mare thickness estimates  
507 would provide a lower bound.

508 On Earth, radial dike swarms have been used to characterise local stress per-

509 turbations induced either by the magmatic plumbing system (Odé, 1957; Muller  
510 and Pollard, 1977; Mériaux and Lister, 2002) or the volcanic edifice (Roman and  
511 Jaupart, 2014). Lessons learned from Earth applied volcanology have been trans-  
512 ferred to planetology with success. In particular, similar giant dike swarms ob-  
513 served on Venus have been interpreted either in terms of local plumbing system  
514 or large scale rising diapir effects (Ernst et al., 1995; Galgana et al., 2013). More  
515 recent studies of magma transport on Earth have relied not only on the magmatic  
516 intrusion shape and orientation but also on vent location to gain quantitative infor-  
517 mation on the stress state, otherwise difficult to constrain, within volcanic edifices  
518 (Maccaferri et al., 2017) or rifting areas (Maccaferri et al., 2014). In particular, the  
519 surface unloading effect induced by caldera formation has been proven to strongly  
520 affects the magma trajectory favouring horizontal deflection at shallow depth and  
521 sill-shaped storage zone formation (Corbi et al., 2015). The present study shows  
522 that applying these new concepts, based on magma ability to reach the surface and  
523 on the path followed by magma in the crust, to planetary volcanism, opens new  
524 perspectives. Additional information on the lithospheric elastic thickness (see  
525 Michaut and Pinel (2018) as well as this study) and on its state of stress can be de-  
526 rived. Conversely, applications of magma transport models on various terrestrial  
527 planets and settings improve their validation to larger parameters ranges.

## 528 **Acknowledgments**

529 This work has been realised thanks to the financial support of the IDEXLyon  
530 Project of the University of Lyon in the frame of the "Programme Investissements  
531 d'Avenir" (ANR-16-IDEX-0005), PNP/INSU/CNES, as well as the Région Au-  
532 vergne Rhône-Alpes. The collaboration between ISTerre and GFZ was funded

533 by the Campus France-DAAD exchange program and the ANR-DFG NLE 2018  
 534 MagmaPropagator projet (ANR-18-CE92-0037). The authors thank the editor  
 535 W. McKinnon and two anonymous reviewers for their help in improving our  
 536 manuscript.

### 537 **Appendix A. Stress due to a cylindrical load on an elastic shell**

538 We use the analytical solution by Brotchie (1971), also used by Solomon and  
 539 Head (1979) and Freed et al. (2001) to calculate the stress in a spherical elastic  
 540 shell of thickness  $T_e$ , Young's modulus  $E$ , Poisson's ratio  $\nu$  due to a cylindrical  
 541 load of radius  $\tau_M$ , thickness  $h_M$ , density  $\rho_M$ . The shell overlies a fluid of density  
 542  $\rho_L$ . We note  $q_M = \rho_M g h_M$  and the flexural rigidity  $D = ET_e^3/(12(1 - \nu^2))$ . We  
 543 note the radius of "relative stiffness"  $L$

$$L = \left( \frac{D}{ET_e/R^2 + \rho_L g} \right)^{1/4} \quad (\text{A.1})$$

544 We also note  $d_M = \tau_M/L$  and  $x = \tau_M + x_M$ , and  $R_s$  the spherical radius to  
 545 the midplane of the shell. Outside the load ( $x_M > 0$ ), the vertical displacement  
 546  $w$ , positive downwards, the radial and azimuthal moment  $M_\phi$  and  $M_\theta$  and radial

547 and azimuthal resultant  $N_\Phi$  and  $N_\Theta$  are :

$$w = \frac{q_M d_M}{ET_e/R_s^2 + \rho_L g} (\text{Ber}' d_M \text{Ker } x - \text{Bei}' d_M \text{Kei } x) \quad (\text{A.2})$$

$$M_\Phi = -q_M d_M L^2 \left[ \text{Ber}' d_M \left( \text{Kei } x + \frac{1-\nu}{x} \text{Ker}' x \right) + \text{Bei}' d_M \left( \text{Ker } x - \frac{1-\nu}{x} \text{Kei}' x \right) \right] \quad (\text{A.3})$$

$$M_\Theta = -q_M d_M L^2 \left[ \text{Ber}' d_M \left( \nu \text{Kei } x - \frac{1-\nu}{x} \text{Ker}' x \right) + \text{Bei}' d_M \left( \nu \text{Ker } x + \frac{1-\nu}{x} \text{Kei}' x \right) \right] \quad (\text{A.4})$$

$$N_\Phi = \frac{q_M ET_e/R_s}{ET_e/R_s^2 + \rho_L g} \left[ \frac{1}{2} \frac{d_M^2}{x^2} + \frac{d_M}{x} (\text{Ber}' d_M \text{Kei}' x + \text{Bei}' d_M \text{Ker}' x) \right] \quad (\text{A.5})$$

$$N_\Theta = -N_\Phi + ET_e w / R_s \quad (\text{A.6})$$

548 where Ber, Bei, Ker and Kei are Bessel-Kelvin functions of order zero and  
549 the prime denotes their first derivatives.

550 Within the shell at  $(R_s - T_e/2) \leq r \leq (R_s + T_e/2)$ , the net horizontal radial  
551 stress  $\sigma_\Phi$  and azimuthal stress  $\sigma_\Theta$  with respect to lead center are:

$$\sigma_\Phi = -\frac{N_\Phi}{T_e} + \frac{12M_\Phi}{Te^3}(r - R_s) \quad (\text{A.7})$$

$$\sigma_\Theta = -\frac{N_\Theta}{T_e} + \frac{12M_\Theta}{Te^3}(r - R_s) \quad (\text{A.8})$$

## 552 References

553 Anderson, E. M., 1951. The dynamics of faulting and dyke formation with appli-  
554 cations to Brittan, Edinburgh. London, Oliver and Boyd.

555 Brotchie, J. F., 1971. Flexure of a liquid-filled spherical shell in a radial gravity  
556 field. Modern Geol. 3, 15–23.

557 Cartwright, J., Moller Hansen, D., 2006. Magma transport through the crust via  
558 interconnected sill complexes. *Geol. Soc. Am. Bull.* 34, 929–932.

559 Corbi, F., Rivalta, E., Pinel, V., Maccaferri, F., Acocella, V., 2016. Understanding  
560 the link between circumferential dikes and eruptive fissures around calderas  
561 based on numerical and analog models. *Geophysical Research Letters* 43 (12),  
562 6,212–6,219.

563 Corbi, F., Rivalta, E., Pinel, V., Maccaferri, F., Bagnardi, M., Acocella, V., 2015.  
564 How caldera collapse shapes the shallow emplacement and transfer of magma  
565 in active volcanoes. *Earth and Planetary Science Letters* 431, 287 – 293.

566 De Hon, R. A., 1979. Thickness of western mare basalts. *Proc. Lunar Planet. Sci.*  
567 *Conf.* 10, 2,935–2,955.

568 Ernst, R., Head, J. W., Parfitt, E., Grosfils, E., 1995. Giant radiating dyke swarms  
569 on Earth and Venus. *Earth Sci. Rev.* 39, 1–58.

570 Evans, R., Soderblom, J. M., Andrews-Hanna, J. C., Solomon, S. C., Zuber, M. T.,  
571 2016. Identification of buried lunar impact craters from GRAIL data and impli-  
572 cations for the nearside maria. *Geophys. Res. Lett.* 43, 2,445–2,455.

573 Freed, A. M., Melosh, H. J., Solomon, S. C., 2001. Tectonics of mascon loading:  
574 Resolution of the strike-slip faulting paradox. *J. Geophys. Res. (Planets)* 106,  
575 20,603–20,620.

576 Galgana, G. A., Grosfils, E. B., McGovern, P. J., 2013. Radial dike formation  
577 on Venus: Insights from models of uplift, flexure and magmatism. *Icarus* 225,  
578 538–547.

- 579 Gong, S., Wieczorek, M. A., Nimmo, F., Kiefer, W. S., Head, J. W., Huang, C.,  
580 Smith, D. E., Zuber, M. T., 2016. Thicknesses of mare basalts on the Moon  
581 from gravity and topography. *J. Geophys. Res. (Planets)* 121, 854–870.
- 582 Jozwiak, L. M., Head, J. W., Wilson, L., 2015. Lunar floor-fractured craters as  
583 magmatic intrusions: Geometry, modes of emplacement, associated tectonic  
584 and volcanic features, and implications for gravity anomalies. *Icarus* 248, 424–  
585 447.
- 586 Jozwiak, L. M., Head, J. W., Zuber, M. T., Smith, D. E., Neumann, G. A., 2012.  
587 Lunar floor-fractured craters: Classification, distribution, origin and implica-  
588 tions for magmatism and shallow crustal structure. *J. Geophys. Res. (Planets)*  
589 117, 10.1029/2012JE004134.
- 590 Laneuville, M., Wieczorek, M. A., Breuer, D., Tosi, N., 2013. Asymmetric ther-  
591 mal evolution of the Moon. *J. Geophys. Res.* 118, 1,435–1,452.
- 592 Maccaferri, F., Bonafede, M., Rivalta, E., 2011. A quantitative study of the mech-  
593 anisms governing dike propagation, dike arrest and sill formation. *J. Volcanol.*  
594 *Geotherm. Res.* 208, 39–50.
- 595 Maccaferri, F., Richeter, N., Walter, T. R., 2017. The effect of giant lateral col-  
596 lapses on magma pathways and the location of volcanism. *Nature Communica-*  
597 *tions* 8, 10.1038/S41467-017-01256-2.
- 598 Maccaferri, F., Rivalta, E., Keir, D., Acocella, V., 2014. Off-rift volcanism in rift  
599 zones determined by crustal unloading. *Nature Geoscience* 7, 297–300.
- 600 McGovern, P. J., Litherland, M. M., 2011. Lithospheric stress and basaltic magma

601 ascent on the Moon, with implications for large volcanic provinces and edifices.  
602 42nd LPSC Abstract, 2,587.

603 Melosh, J., 1976. On the origin of fractures radial to lunar basins. Proc. Lunar Sci.  
604 Conf. 7th, 2,967–2,982.

605 Menand, T., 2011. Physical controls and depth of emplacement of igneous bodies:  
606 A review. *Tectonophysics* 500, 11–19.

607 Menand, T., Daniels, K. A., Benghiat, P., 2010. Dyke propagation and sill  
608 formation in a compressive tectonic environment. *J. Geophys. Res.* 115,  
609 10.1029/2009JB006791.

610 Mériaux, C., Lister, J. R., 2002. Calculation of dike trajectories from volcanic  
611 centres. *J. Geophys. Res.* 107(B4), 1–10.

612 Michaut, C., Pinel, V., 2018. Magma ascent and eruption triggered by cratering  
613 on the Moon. *Geophys. Res. Lett.* 45, 10.1029/2018GL078150.

614 Michaut, C., Thiriet, M., Thorey, C., 2016. Insights into mare basalt thicknesses  
615 on the Moon from intrusive magmatism. *Phys. Earth Planet. Int.* 257, 187–192.

616 Muirhead, J. D., Airoidi, G., Rowland, J. V., Connor, L., White, J. D. L., 2012.  
617 Interconnected sills and inclined sheet intrusions control shallow magma trans-  
618 port in the Ferrar large igneous province, Antarctica. *Geol. Soc. Am. Bull.* 124,  
619 162–180.

620 Muller, O. H., Pollard, D. D., 1977. The stress state near Spanisk Peaks, Colorado,  
621 determined from a dike pattern. *Pure Appl. Geophys.* 115, 69–86.

- 622 Nakamura, K., 1977. Volcanoes as possible indicators of tectonic stress orienta-  
623 tion - principle and proposal. *J. Volc. Geotherm. Res.* 2, 1–16.
- 624 Neumann, G. A., Zuber, M. T., Wieczorek, M. A., McGovern, P. J., Baker, D.  
625 M. H., Solomon, S. C., Smith, D. E., Lemoine, F. G., Mazarico, E., Sabaka,  
626 T. J., Goossens, S. J., Melosh, H. J., Phillips, R. J., Asmar, S. W., Kono-  
627 pliv, A. S., Williams, J. G., Sori, M. M., Soderblom, J. M., Miljkovic, K.,  
628 Andrews-Hanna, J. C., Nimmo, F., Kiefer, W. S., 2015. Lunar impact basins  
629 revealed by gravity recovery and interior laboratory measurements. *Sci. Adv.* 1,  
630 10.1126/sciadv.1500852.
- 631 Odé, H., 1957. Mechanical analysis of the dike pattern of the Spanish Peaks area,  
632 Colorado. *Geol. Soc. Am. Bull.* 68, 567–576.
- 633 Peeples, W. J., Sill, W. R., May, T. W., Ward, S. H., Philips, R. J., Jordan, R. L.,  
634 Abbott, E. A., Killpack, T. J., 1978. Orbital radar evidence for lunar subsurface  
635 layering in Maria Serenitatis and Crisium. *J. Geophys. Res.* 83, 3,459–3,468.
- 636 Pinel, V., Carrara, A., Maccaferri, F., Rivalta, E., Corbi, F., 2017. A two-step  
637 model for dynamical dike propagation in two-dimensions: Application to  
638 the July 2001 Etna eruption. *Journal of Geophysical Research: Solid Earth*,  
639 10.1002/2016JB013630.
- 640 Pinel, V., Jaupart, C., 2000. The effect of edifice load on magma ascent beneath a  
641 volcano. *Phil. Trans. Roy. Soc. Lond.* 358, 1,515–1,532.
- 642 Pritchard, M., Stevenson, D. J., 2000. Thermal aspects of lunar origin. In: *Origin*  
643 *of the Earth and Moon*. Tucson, Univ. of Arizona Press, pp. 179–196.



- 644 Richardson, J. A., Connor, C., Wetmore, P., Connor, L., Gallant, E., 2015. Role of  
645 sills in the development of volcanic fields: Insights from lidar mapping surveys  
646 of the San Rafael Swell, Utah. *Geology* 43, 10.1130/G37094.1.
- 647 Roman, A., Jaupart, C., 2014. The impact of a volcanic edifice on intrusive and  
648 eruptive activity. *Earth Planet. Sc. Lett.* 408, 1–8.
- 649 Schultz, P. H., 1976a. Floor-fractured lunar craters. *Moon* 15, 241–273.
- 650 Schultz, P. H., 1976b. Moon morphology: Interpretations based on lunar orbiter  
651 photography. University of Texas Press, Austin, Texas, p. 626.
- 652 Sneddon, I. N., 1951. Fourier Transforms. Dover Publications, New York.
- 653 Solomon, S. C., Head, J. W., 1979. Vertical movement in mare basins: Relation to  
654 mare emplacement, basin tectonics, and lunar thermal history. *J. Geophys. Res.*  
655 (Planets) 84, 1,667–1,682.
- 656 Thomas, R. J., Rothery, D. A., Conway, S. J., Anand, M., 2015. Explosive volcan-  
657 ism in complex impact craters on Mercury and the Moon: Influence of tectonic  
658 regime on depth of magmatic intrusion. *Earth Planet. Sc. Lett.* 431, 164–172.
- 659 Thorey, C., Michaut, C., 2014. A model for the dynamics of crater-centered in-  
660 trusions: application to lunar floor-fractured craters. *J. Geophys. Res. (Planets)*  
661 119, 286–312.
- 662 Thorey, C., Michaut, C., Wieczorek, M., 2015. Gravitational signatures of lunar  
663 floor-fractured craters. *Earth Planet. Sc. Lett.* 424, 269–279.
- 664 Trang, D., Gillis-Davis, J. J., Hawke, B. R., 2016. The origin of lunar concentric  
665 craters. *Icarus* 278, 62–78.

- 666 Watanabe, T., Masuyama, T., Nagaoka, K., Tahara, T., 2002. Analog experiments  
667 on magma-filled cracks: Competition between external stresses and internal  
668 pressure. *Earth, Planets, Space* 54, 1247–1261.
- 669 Weider, S. Z., Crawford, I. A., Joy, K. H., 2010. Individual lava flow thicknesses in  
670 Oceanus Procellarum and Mare Serenitatis determined from Clementine multi-  
671 spectral data. *Icarus* 209, 323–336.
- 672 Wichman, R. W., Schultz, P. H., 1995. Floor-fractured craters in Mare Smythii  
673 and West of Oceanus Procellarum implications of crater modification by vis-  
674 cous relaxation and igneous intrusion models. *J. Geophys. Res. (Planets)* 100,  
675 21,209–21,218.
- 676 Wieczorek, M. A., Neumann, G. A., Nimmo, F., Kiefer, W. S., Taylor, G. J.,  
677 Melosh, H. J., Phillips, R. J., Solomon, S. C., Andrews-Hanna, J. C., Asmar,  
678 S. W., Konopliv, A. S., Lemoine, F. G., Smith, D. E., Watkins, M. M., Williams,  
679 J. G., Zuber, M. T., 2013. The crust of the Moon as seen by GRAIL. *Science*  
680 339, 671–675.
- 681 Wilson, L., Head, J. W., 2017. Generation, ascent and eruption of magma on the  
682 Moon: New insights into source depths, magma supply, intrusions and effu-  
683 sive/explosive eruptions (Part 1: Theory). *Icarus* 283, 146–175.
- 684 Wood, J. A., Dickey, J. S., Marvin, U. B., Powell, B. N., 1970. Lunar anorthosites  
685 and a geophysical model of the Moon. *Proc. Apollo 11 Lunar Sci. Conf.*, 1965–  
686 1988.

