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# A giant volcanic island in an early Martian Ocean?

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2

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25 **Abstract**

26 Tracking early oceans on Mars has fundamental climatic and exobiological implications  
27 but requires unambiguous datable markers. Here we show that the Olympus Mons giant volcano  
28 shares morphological similarities with active volcanic islands on Earth where major  
29 constructional slope breaks systematically occur at the sea-air transition in response to sharp lava  
30 viscosity contrasts. We propose that the upper rim of the 6-km high concentric main escarpment  
31 surrounding Olympus Mons most likely formed by lava flowing into liquid water when the  
32 edifice was an active volcanic island during the late Noachian - early Hesperian. Similar features  
33 on the northern flank of the 1800 km-distant Alba Mons volcano further support an extended  
34 early Ocean initially occupying Martian lowlands. Abnormal current shoreline heights imply  
35 major surface uplift promoted by internal dynamics during the main phase of development of the  
36 Tharsis Bulge at some time during the Hesperian. After ocean retreat, the uplifted shorelines were  
37 partly blanketed by late sub-aerial activity during the Amazonian. The newly proposed volcanic  
38 shorelines can be dated by radiometric methods, constituting exceptional targets to track early  
39 liquid water at key temporal steps and provide valuable insights into the timing and fate of  
40 Martian Oceans.

41

42 **Keywords:** *Olympus Mons, Volcanic Island, Basal Escarpment, sea-air transition, Subsidence, Uplift*

43

## 44 **1. Introduction**

45           Increasing evidence support the existence of early ocean(s) of liquid water at the surface  
46 of Mars until at least 3.0 Ga ago (Schmidt et al., 2022). Where and when such early ocean(s)  
47 were has fundamental implications regarding Mars climatic evolution and potential extra-  
48 terrestrial life, but remains poorly constrained. In Martian northern plains, two main sets of paleo-  
49 shorelines have been attributed to two independent bodies of liquid water that were separated  
50 from each other by hundreds of millions of years, between ca 3.8 and 3.0 Ga (e.g. Sholes et al.,  
51 2021). Nevertheless, their elevations vary by several kilometres, which may either reflect large  
52 fluctuations in ocean(s) level and/or significant ups and downs of the planet surface in response  
53 to deformation associated with internal dynamics (e.g. Citron et al., 2018; Broquet and Andrews-  
54 Hanna, 2022). Improved knowledge of sea level at key steps of planetary evolution is therefore  
55 needed to better constrain ocean(s) reconstructions and elucidate the fate of early liquid water at  
56 the surface of Mars. Martian volcanoes constitute potential targets of major interest for such  
57 purpose, because (1) their formation under aerial/submarine conditions may have produced  
58 distinctive morphologies, and (2) radiometric dating is achievable on most volcanic products,  
59 potentially providing key absolute ages in the future. Here we identify a new possible marker of  
60 an ancient shoreline, the fossil passage zone of sub-aerial lavas flowing into a liquid ocean, which  
61 seems to be well preserved at the upper rim of the Olympus Mons basal escarpment. The  
62 apparently excessive height of this new marker can be explained by high-magnitude surface uplift  
63 promoted by large-scale mantle upwelling, as shown by new modelling of vertical movements.

64           Olympus Mons is a more than 20 km-tall shield volcano located on the northwestern slope  
65 of the Tharsis rise (Fig. 1). This giant volcanic edifice ( $>300000 \text{ km}^2$ ) was most likely formed by  
66 protracted hot-spot volcanism above a stationary lithosphere in the absence of plate tectonics.  
67 Reported ages for Olympus Mons volcanic activity range from 3.8 Ga to less than 10 Ma (Hiller  
68 et al., 1982; Neukum et al., 2004; Werner, 2009; Isherwood et al., 2013). Most of the volcano  
69 growth is believed to have taken place before  $2.54^{+0.55}_{-0.69}$  Ga (Isherwood et al., 2013), which

70 comprises the admitted period for a possible early Martian Ocean prior to 3.0 Ga. Olympus Mons  
71 exhibits striking morphological features from the centre to the outer parts: (1) The summit area,  
72 which includes several nested elliptic depressions formed by vertical caldera collapse (e.g.  
73 Mouginis-Mark, 2018); (2) a series of morphological “terraces”, which may either mark the edge  
74 of successive thick lava flows or centripetal thrusts formed during gravitational spreading of the  
75 volcano (Musiol et al., 2016); (3) arcuate cliffs, referred to as the “basal escarpment”, which can  
76 be traced all around the volcano; (4) a large “aureole” of chaotic deposits, which may represent  
77 either debris-avalanches deposits formed by flank collapse processes, highly fragmented lava  
78 flows, or late moraine-like glacial deposits (e.g. Lopes et al., 1980; Head et al., 2005); and (5)  
79 recent volcano-tectonic features including narrow grabens and vents, which developed in various  
80 sectors of the volcano during the late Amazonian (Peters and Christensen, 2017).

81         The origin of the basal escarpment remains enigmatic. Part of it has been interpreted as a  
82 constructional feature possibly influenced by the presence of liquid water (De Blasio, 2012). Yet,  
83 the exact relationships between volcano growth and potential water height/depth remain elusive,  
84 as the upper tip of most of the escarpment is several kilometres higher than any reported sea level  
85 for a past Martian Ocean (e.g. Sholes et al., 2021). Alternatively, the escarpment has been widely  
86 interpreted as a series of scarps produced by landslide processes, which truncated the distal flank  
87 of the volcano and cumulatively fed the aureole deposits (e.g. Lopes et al., 1980; De Blasio,  
88 2018). However, most of the volcano outline is sub-circular, i.e. concave towards the centre. This  
89 seems incompatible with formation of most of the escarpment by lateral slides, which would  
90 produce opposite scarp curvature (concave outwards).

91

## 92 **2. Morphology of Olympus Mons and characteristics of the escarpments**

93         MOLA topographic data (Smith et al., 1999) allow further identification of striking  
94 morphological characteristics at the full scale of Olympus Mons (Fig.1):

- 95 - The volcanic edifice is asymmetrical. The western flank covers a wider area than the  
96 eastern flank (Fig. 1a), and typically displays lower slopes (Fig. 1b).
- 97 - Two main sub-circular basal escarpments are visible in the SE and NW sectors (C1 and  
98 C2, Fig. 1a). They can be traced over hundreds of kilometres and are noticeably concave  
99 to the centre of the volcano. Both interrupt abruptly the gentle distal flanks of the shield  
100 and show relatively steep slopes, with individual values up to 35-40° (Fig. 1b) and  
101 average escarpment slopes around 14° and 16° along profiles 2 and 6, respectively (Fig  
102 1c).
- 103 - The NW and SE escarpments are ca 6 and 7 km high, respectively, due to a slight  
104 westward increase in basement depth. Their upper tips, however, occur at similar  
105 elevations, thus supporting the inference of a former presence of a common horizontal  
106 free surface all around the volcano.
- 107 - Additional concave cliffs can be traced in the southern sector. They appear less prominent,  
108 being partly smoothed by late lava fields (Fig. 1).
- 109 - In the northern and western sectors, the C2 escarpment is interrupted by convex U-shaped  
110 depressions bordered by debris fields typical of flank collapse processes (e.g. Siebert,  
111 1984; Hildenbrand et al., 2018). Other subtle U-shaped scars in the NW and NE sectors  
112 were also likely formed by lateral flank destabilization, and later smoothed by volcanic  
113 products including a large lava field (Fig. 1 b, c).
- 114 - Many other parts of the volcano have been re-surfaced. The basal escarpments, especially,  
115 have been deeply modified by smaller-scale landslides, canyon incision, accumulation of  
116 piedmont deposits, or cliff concealing by recent lava flows. Nevertheless, these secondary  
117 processes did not erase the major morphological characteristics, especially the regional  
118 expression of the C1 and C2 prominent cliffs.

119

### 120 **3. Comparison with current volcanic ocean islands on Earth**

121 Intra-plate volcanic islands on Earth rise up to 10 km from the oceanic basement and may  
122 constitute our best terrestrial analogues to Olympus Mons. Among many examples in various  
123 provinces (e.g. Azores, Canary, Cape Verde, Reunion, Hawaii, Galapagos), we selected three  
124 islands with contrasted sizes and varied eruptive histories. Both show well-preserved young  
125 constructional morphologies, along with older sectors affected by other geological and coastal  
126 processes (Fig. 2).

127 Pico Island (Azores, Atlantic Ocean), the smallest, was formed by successive episodes of  
128 growth and repeated destabilization by lateral landslides during the last 200 kyr (Hildenbrand et  
129 al., 2012; Costa et al., 2015; Marques et al., 2021). The young Pico Volcano (PV in Fig. 2) is  
130 elliptic in plan view. Topographic cross-sections (Fig. 2) show a steep summit area ( $35^\circ$ )  
131 followed by a gentle regular northern sub-aerial flank ( $<10^\circ$ ). A sharp break occurs close to the  
132 air-sea transition, with a main slope increase by a factor 2, i.e. an angular difference of about  $10^\circ$ .  
133 The PV's southern flank is shorter and truncated by a steep U-shape landslide structure yielding  
134 an inward-convex coastal embayment.

135 Fogo Island (Cape Verde, Atlantic), of intermediate size, comprises a main former  
136 volcanic edifice that was destroyed by a massive east-directed flank collapse (Day et al., 1999)  
137 about 50 ka ago (Marques et al., 2019). The preserved western flank of the former volcano  
138 appears sub-circular in plan view, and shows typical constructional profiles (Fig. 2). A sharp  
139 morphological break occurs at shore level, with sub-aerial and submarine constructional slopes of  
140 ca  $10^\circ$  and  $25^\circ$ , respectively. The still-active Fogo volcano that grew on the landslide scar also  
141 shows a short eastern flank with a pronounced slope break at sea level.

142 The Big Island (Hawaii, Pacific Ocean), by far the largest on Earth, comprises five  
143 coalescent shield volcanoes with ages decreasing towards the southeast (Sherrod, et al., 2021).  
144 The flanks of the Mauna Loa active volcano (Fig. 2) are cut by several large flank collapse  
145 structures partly buried by recent volcanic products (e.g. Moore et al., 1994; Sherrod et al., 2021).



146 The conspicuous Hilina slump is presently displacing the whole SE flank downwards at rates up  
147 to 10 cm/yr (e.g. Morgan et al., 2000; Owen et al., 2000), thus making a significant step (Fig. 2).  
148 Topographic profiles in zones not affected by major slides show typical constructional slopes. A  
149 marked angular difference of about  $13^\circ$  is observed at the air-sea transition, with average values  
150 around  $4^\circ$  and  $17^\circ$  for the sub-aerial and submarine slopes of Mauna Loa, respectively. The  
151 Kilauea active volcano developed at the head of the Hilina slump shows a similar constructional  
152 slope break at shore level. Deeper steps in the eastern and northern sectors likely witness older  
153 shorelines carried down by island subsidence due to loading of the lithosphere during/after the  
154 main shield-building phase (e.g. Taylor, 2019, and references herein).

155 The youngest portions of the three selected volcanic islands exhibit clear constructional  
156 main slope breaks at shore level, by up to  $15^\circ$  in magnitude. These can be explained by sharp  
157 viscosity contrasts due to differential lava cooling into air or water (Griffiths, 2000) as currently  
158 observed at the passage zone between submarine and sub-aerial volcanic series on uplifted  
159 oceanic islands, e.g., Santiago in Cape Verde or Santa Maria in the Azores (Marques et al.,  
160 2020a,b). In the latter case (Fig. 3), the main constructional slope break coincides with a sudden  
161 increase in lava flow dip, from gentle in the sub-aerial part to steeper in the submarine portions,  
162 where pillow-lavas and pillow breccias witness fast cooling in the presence of liquid water (Fig.  
163 3). By analogy, the main slope break (ca.  $13^\circ$ ) at the upper tips of the C1 and C2 inward concave  
164 escarpments surrounding Olympus Mons (Fig. 1) supports contrasted lava viscosity at shore level.  
165 This leads us to propose that Olympus Mons was a former volcanic island surrounded by liquid  
166 water.

167 One may argue that alternative processes like lava-ice interactions should be considered,  
168 as Martian northern lowlands may have been occupied by a (partially?) frozen ocean during the  
169 late Hesperian (Ivanov et al., 2014, 2015). On Earth, lava-ice interactions produce a variety of  
170 glacio-volcanic edifices, referred to as “tuyas” (see review in Russell et al., 2014). Flat-topped

171 tuyas, especially, form as magma rises under a thick ice cap leading to development of a transient  
172 lake. As in marine environments, early interactions between hot lavas and melt water sometimes  
173 produce subaqueous tephra and/or pillow-lavas, followed by the eruption of sub-aerial lavas once  
174 the lake has been drained. As a result, flat-topped “tuya” comprise an upper gentle plateau  
175 surrounded by steep slopes, making a significant morphological step like the main slope break  
176 around Olympus Mons. Nevertheless, tuya formation is a very local process that typically occurs  
177 during a single main eruption, and is therefore an “instantaneous” process at geological  
178 timescales. Terrestrial tuyas have small dimensions (few kilometres width, few hundred meters  
179 high), and rarely exhibit conical regular morphologies. For these reasons, it seems highly  
180 unplausible that the well-behaved slope break observed for hundreds of kilometres around  
181 Olympus Mons could have been formed simply by past interactions between a batch of rising  
182 magma and a thick frozen ocean.

183         Lava flowing into a large ocean of (still) liquid water thus remains the most likely  
184 scenario to explain both the sub-circular aspect of the main escarpments surrounding Olympus  
185 Mons and the conspicuous slope-break with a main angular difference by ca 13°. Conversely, late  
186 lava fields (Fig. 1, profile 3) show monotonous and regular slopes typical of terrestrial sub-aerial  
187 volcanoes. These low-viscosity lavas most likely partially concealed the escarpment cliffs after  
188 water retreat.

189

#### 190 **4. Ups and downs: Olympus Mons vs Martian ocean(s)**

191         The average height of the main escarpments suggests a 6 km-thick column of liquid water  
192 initially surrounding the inferred volcanic island. This is more than the average depth of oceans  
193 on Earth, and twice the plausible depth for an early Martian ocean ca 3.8 Ga ago (Parker et al.,  
194 1993; Sholes et al., 2021). Nevertheless, the escarpment height is not inconsistent with a globally  
195 shallow ocean. Massive loading by volcano construction may, indeed, have favoured the fast

196 development of a peripheral flexural basin, as observed for oceanic lithosphere on Earth (Watts  
197 and Zhong, 2000), leading to increasing water column around the growing edifice.

198         The earliest stages of growth of Olympus Mons remain poorly dated. Ages around 3.8 Ga  
199 were reported from in-situ crater counting (HRSC data) near the upper tip of the northwestern  
200 basal escarpment (Neukum et al., 2004). These ages have been later discarded (Isherwood et al.,  
201 2013), as they were based on only 4 craters of large size ( $>1 \text{ km}^2$ ). Additional crater-counting age  
202 estimates (Hiller et al., 1982; Isherwood et al., 2013) have been proposed, but these were  
203 acquired on indirect, distant markers. An age value of  $3.67^{+0.05}_{-0.10}$  Ga (Isherwood et al., 2013) was  
204 derived for a “discordant” lava flow exposed 500 km away from the volcano’s NE outline.  
205 Deviation in the path of the flow from the down-slope direction of the present-day flexural trough  
206 is believed to indicate that the flow pre-dates downward lithosphere flexure by main volcano load,  
207 and therefore that the dominant growth of Olympus Mons occurred after  $3.67^{+0.05}_{-0.10}$  Ga  
208 (Isherwood et al., 2013). However, this flow is located tens of kilometers away from the main  
209 though rim, and its sinuous path might just reflect channeling of the lava in a local topography  
210 either before, during, or after lithosphere flexure. On the other hand, the northern aureole deposits  
211 have been attributed to late landslide processes post-dating Olympus Mons main building-phase.  
212 Crater-counting in the northern aureole yielded initial age estimates between 3.68 Ga and 3.51 Ga  
213 (Hiller et al., 1982). More recent estimates (Isherwood et al., 2013) range between  $3.47^{+0.07}_{-0.16}$  Ga  
214 and  $0.98^{+0.51}_{-0.61}$  Ga, with a preferred value of  $2.54^{+0.55}_{-0.69}$  Ga, i.e up to 2.99 Ga when uncertainties  
215 are accounted for. Therefore, the bulk of the volcano most likely grew up prior to 3.0 Ga, and  
216 possibly as early as 3.7-3.6 Ga ago if we consider the oldest ages reported for both the basal  
217 escarpment and the northern aureole deposits (Hiller et al., 1982; Neukum et al., 2004; Werner,  
218 2009; Isherwood et al., 2013).

219         Noticeably, the northern aureole was partly fed by the recent slide interrupting the C2  
220 concave cliff here interpreted as a former shoreline (Fig. 1). Olympus Mons thus most likely was

221 already a large and thick volcanic island with dimensions comparable to present-day when the  
222 slide occurred. Significant flexural subsidence and basin deepening by dominant volcano load  
223 may thus have taken place during either the late Noachian or early Hesperian (Fig. 4a, b), which  
224 is significantly earlier than previously proposed (Isherwood et al., 2013).

225         Figure S1 shows our new modelling results for downward loading of the lithosphere by a  
226 former edifice with variable thicknesses (model description in Appendix A). An initial volcano  
227 height of 15-20 km is sufficient to produce a 1300 km-wide flexural basin (Fig. S2) and locally  
228 increase global ocean water depth by a few kilometres around the volcano. Nevertheless, the  
229 basal escarpment upper tips stand more than 6.5 km higher than the values commonly reported  
230 for early shorelines on Mars (Arabia level, 3.7 Ga, elevation between -3.6 km (Sholes et al.,  
231 2021) and -1.6 km (Parker et al., 1993)), and about 10 km higher than the youngest ones  
232 (Deuteronilus level, ca 3.0 Ga, mean elevation around - 4 km (Sholes et al., 2021; Schmidt et al.,  
233 2022)). Therefore, generalized island uplift by at least 6 - 7 km is needed to explain the present  
234 position of the main slope break.

235         Dynamic swelling (plume load) associated with main development of the Tharsis Rise  
236 during the early to middle Hesperian (Bouley et al., 2018) could have promoted surface uplift by  
237 several hundreds of meters to kilometres (Fig. 4c), as recorded by some oceanic volcanoes on  
238 Earth (e.g. Moore et al., 1999; Zhong and Watts, 2002; Hildenbrand et al., 2003; Marques et al.,  
239 2020a,b). Dynamic uplift is here tested considering upward impact of the Martian lithosphere by  
240 an elliptical mantle anomaly centered underneath OM (Fig. S3). A plume with a 1000 km radius  
241 and 650 km thickness, an excess temperature of 300 °C and thus a buoyancy of 1% with respect  
242 to ambient mantle can produce a swell exceeding 6 km. Possible effects of magma underplating  
243 are further tested considering low-density material ponded underneath the Martian crust (Fig. S4).  
244 A 10 km-thick body with a geometry mirroring volcano topography could produce additional  
245 surface rise of more than 2 km. Other mechanisms such as multiple sill intrusions during recent  
246 volcanic activity (Fig. 4d) or late isostatic readjustment after melting of a thick icesheet could

247 also contribute to further edifice uplift but may be rather limited, as inferred from glacial isostatic  
248 adjustment on Earth (e.g. Steffen and Wu, 2011).

249

## 250 **5. A scenario applicable to other Martian volcanoes?**

251 Alba Mons appears a good candidate to further track the presence of past liquid water on  
252 Mars and evaluate vertical movements at regional scales. Alba Mons straddles the northwestern  
253 edge of the Tharsis rise, 1800 km North of Olympus Mons (Fig. 5). Estimated ages from crater  
254 counting suggest dominant volcano construction between 3.6 Ga and 3.2 Ga (Ivanov and Head,  
255 2006), which partly overlaps the updated time pace for a possible Martian ocean (Schmidt et al.,  
256 2022) and the currently inferred age for the development of the Tharsis bulge (Bouley et al.,  
257 2016; Citron et al., 2017; Broquet and Andrews-Hanna, 2022). The Alba Mons asymmetric  
258 shield-volcano is outlined by arcuate volcano-tectonic structures and sharp N-S lineaments  
259 associated with late-stage faulting (e.g. Ivanov and Head, 2006; Leone et al., 2021).

260 While most of the edifice features very gentle regular flanks, the northern edge shows a  
261 clear slope break (Fig. 5). The observed angular difference ( $<2^\circ$ ) is much smaller than in the case  
262 of Olympus Mons and terrestrial cases. This possibly indicates lower viscosity of the magma  
263 erupted at Alba Mons, or partial concealing/smoothing of an initial steeper slope break by late  
264 lavas, as suspected for some sectors of Olympus Mons (Figure 1, profile 7). Nevertheless, the  
265 relative jump in slope by a factor 2 across the northern flank of Alba Mons is still significant, and  
266 may witness a former passage zone from sub-aerial to submarine environments, as proposed for  
267 the C1 and C2 cliffs around Olympus Mons. The smaller height of Alba Mons relatively steep  
268 lower flank ( $<5$  km) may reflect (a) the peculiar position of the edifice at the margin of an ocean  
269 basin, inducing a shallower water depth, (b) reduced flexural subsidence due to a shorter volcanic  
270 edifice ( $<8$  km), (c) a younger and globally shallower ocean during the early Hesperian, and (d)

271 emplacement of the volcanic edifice during/after initiation of the uplift associated with the  
272 Tharsis Rise.

273 Despite significant differences, Alba Mons and Olympus Mons overall share common  
274 morphological characteristics, which support two main stages of evolution in the northwest  
275 Tharsis region: (1) early volcanic construction in the presence of liquid water in the northern  
276 plains, and (2) significant regional uplift during early-middle Hesperian, eventually followed by  
277 late sub-aerial volcanism after water retreat.

278

## 279 **6. Broader implications**

280 The new markers of sea level here proposed stand much higher than elevated shorelines  
281 previously identified around the Tharsis Rise (longitudes between  $-150^{\circ}\text{E}$  and  $-120^{\circ}\text{E}$ ),  
282 supporting significant additional uplift. From our new modelling, a buoyant mantle plume with a  
283 1000 km radius centered underneath Olympus would produce no significant surface uplift at Alba  
284 Mons. A larger plume (2000 km radius) impacting the lithosphere below the central part of the  
285 Tharsis bulge ca 3.5 Ga ago could have produced comparable large magnitude surface uplift of  
286 both volcanoes. However, such large-scale doming would cause differential uplift between the  
287 western and eastern parts of Olympus Mons, which is not supported by the comparable elevation  
288 of the C1 and C2 upper escarpment rims. Alternatively, both volcanoes may have experienced  
289 regional uplift in response to upward loading by distinct plumes of moderate sizes. Such features  
290 may be connected to a wider deep mantle anomaly under the Tharsis region, as proposed for the  
291 South Pacific Superswell on Earth, where plumelets separated by hundreds of km rise  
292 periodically from a super-plume blocked at depth (Courillot et al., 2003; Bonneville et al., 2006).  
293 Rejuvenated volcanism ( $<3$  Ga) on Olympus Mons and at the scale of the whole Tharsis dome  
294 (e.g. Leone et al., 2021) further suggests extended and protracted partial melting, supporting  
295 large-scale mantle upwelling during the late Hesperian and part of the Amazonian. According to

296 a recent study (Broquet and Andrews-Hanna, 2022), a currently still-active large plume head  
297 (nominal radius around 2000 km) could account for a surface uplift of about 1 km for the  
298 Elysium Planitia region during the late Amazonian, further supporting the protracted influence of  
299 deep Martian geodynamics as a main carrier of surface deformation.

300

## 301 **7. Conclusions**

302 Conspicuous slope breaks at the top of the main basal escarpment surrounding Olympus  
303 Mons are here interpreted as paleo-shorelines around a former volcanic island. Similar features at  
304 Alba Mons further support the existence of an early ocean occupying the northern plains of Mars  
305 along the NW edge of the Tharsis region. The proposed shorelines stand much higher than  
306 previously reported late Noachian sea levels, suggesting later high-magnitude regional uplift in  
307 response to deep internal forcing. Plume buoyancy leading to upward deformation of the Martian  
308 lithosphere associated with the main development of the Tharsis Rise could adequately explain  
309 such large-scale vertical movements followed by widespread late sub-aerial volcanism in the  
310 region.

311 Improved age determination is needed to better constrain the timing for ocean stability  
312 and the dynamics of large vertical movements. Volcanic units involved in the slope breaks at  
313 Olympus Mons constitute exceptional targets for such a purpose. Recent analytical advances in  
314 in-situ K-Ar radiometric dating (e.g Cho and Cohen, 2018; Cattani et al., 2019; Cohen et al.,  
315 2021) presently allow age measurement on Mars-like volcanic materials with relative precision of  
316 only a few percent. Future spacecrafts dedicated to sample return and/or rover equipped for in-  
317 situ dating on selected sites of Olympus Mons constitute a promising line of research for the  
318 future, which can have significant impact regarding the longevity of oceans and the potential fate  
319 of early life on Mars.

320

321 **Appendix A: Methods**

322 Shaded-relief maps, slope maps, and topographic profiles of Olympus and Alba Mons  
323 have been generated from the MOLA global digital elevation model<sup>16</sup> with a spatial resolution of  
324 200 m. The size of each DEM cell is very small compared to the actual size of both volcanoes (>  
325 300000 km<sup>2</sup> each), enabling a fine assessment of the main morphological characteristics of both  
326 edifices. Morphological analyses of the three volcanic islands here considered have been  
327 achieved from DEMs merging topographic and bathymetric data. Source of the data for Pico is  
328 from the EMODNET European open-access database. Public data for Hawaii  
329 (<http://www.soest.hawaii.edu/hmrg/multibeam/bathymetry.php>) is available from the School of  
330 Ocean and Earth Science and Technology (SOEST, University of Hawaii). Data for Fogo (Le Bas  
331 et al., 2007) have been made available to us by T. Le Bas (Univ. of Southampton, UK).

332 Modelling the deformation induced by either volcano load, upward plume load, and  
333 magma underplating has been achieved considering a purely elastic behaviour. Proper estimation  
334 of the lithosphere elastic thickness at the time of loading is essential for such purpose. Seismic  
335 data acquired recently during the insight mission in Elysium Planitia about 5 000 km away from  
336 Olympus Mons have shown the presence of deep reflectors interpreted as evidence for current  
337 thickness of 500 km for the Martian lithosphere (Kahn et al., 2021). However, a much thinner  
338 lithosphere thickness is expected at the time of loading here considered, mostly between ca 3.8  
339 and 3.0 Ga ago.

340 Bending of an elastic plate depends on the acting forces and the elastic parameters of the  
341 plate. For a two-dimensional periodic loading and laterally constant elastic parameters, the  
342 following formula describes bending (Turcotte and Schubert, 2002):

343 
$$w(x) = \frac{F_l}{F_r + D \left(\frac{2\pi}{\lambda}\right)^4} \quad (\text{eq. 1})$$

344



345 where  $w$  [m] is the deformation of the plate in the (vertical) direction of the applied force,  $F_1 =$   
 346  $\rho_r * g * h$  [N/m<sup>2</sup>] is the loading pressure with  $\rho_r$  the effective density of the restoring mass and  $h$  the  
 347 height of the load,  $F_r$  [N/m<sup>3</sup>] is the restoring pressure per unit length of deformation  $w$ ,  $D =$   
 348  $ET_e^3/[12(1-\nu^2)]$  is the flexural rigidity [Nm] with  $E$  [Pa] Young's modulus,  $\nu$  Poisson's ratio and  
 349  $T_e$  [m] the equivalent elastic thickness of the deformed plate.  $\lambda$  [m] is the wavelength of the  
 350 periodic load, the load having thus the form of  $F_1 = \rho_l * g * h_0 * \sin(2\pi/\lambda * x)$  where  $\rho_l$  [kg/m<sup>3</sup>] is the  
 351 effective density of the load,  $g$  [m/s<sup>2</sup>] the gravity acceleration and  $h_0$  [m] the amplitude of load  
 352 height (topography or plume thickness). The actual distribution of the forces is evidently not  
 353 periodic, but it is possible to apply eq. 1 to the spectral components of the force distribution. Thus,  
 354 topography and plume profiles are analysed using Fast Fourier Transform, each coefficient of the  
 355 resulting spectrum is multiplied by eq. 1 using the wavelength corresponding the respective  
 356 coefficient and the filtered spectrum is then transformed back into real space.

357 For a topographic load, the density  $\rho_l$  is the density of the material building the  
 358 topography whereas the effective density of the restoring mass is  $\rho_r = \rho_m - \rho_f$ ,  $\rho_m$  being the  
 359 density of the mantle and  $\rho_f$  the density of the infill of the flexural basin. Since we suppose that  
 360 the basin was filled with water, we use for  $\rho_f$  1000 kg/m<sup>3</sup>. For a plume load, the effective density  
 361 of the load is the density difference between the hot plume and the surrounding mantle:  $\rho_l = \rho_m -$   
 362  $\rho_p$ .  $\rho_p$  depends on the temperature difference  $\Delta T$  [K]:  $\rho_p = \rho_m * (1 - k\Delta T)$ ,  $k$  [1/K] being the  
 363 thermal expansion coefficient, taken as  $3.5 * 10^{-5} \text{ K}^{-1}$  (e.g. Afonso et al., 2005 for terrestrial  
 364 mantle). The effective density is thus  $\rho_l = -\rho_m * k\Delta T$ . For the restoring force,  $\rho_r$  is the density of  
 365 the crust  $\rho_c$  for which we take the same density as for the volcanic material.

366 In the case of Olympus Mons, the values used for the calculations are summarized in  
 367 Table 1. They were mostly taken from Belleguic et al. (2005). This model scales with the model  
 368 from Broquet and Andrews-Hanna (2022). Figure S1 shows results for the calculation of

369 topography resulting from initial volcanism on a plate of variable equivalent elastic thickness  
370 (EET). The thickness of deposited volcanic material was adjusted for each tested EET such that  
371 the break in slope is located at sea level (defined as 0 m topography). From a point of view of  
372 topography of the main volcanic building, EETs from 30 to 150 km are possible if loading has  
373 thicknesses between 40 and 18 km respectively. However, the different models are clearly  
374 distinguished by the distance of the bulge from the volcanic centre, which varies from 400 to  
375 1000 km. Fig. A2 shows that the bulge visible on the topographic map of OM has a diameter of  
376 about 1300 km. An EET of 80 to 100 km with an initial load of about 20 km thickness is thus the  
377 most probable scenario during the initial build-up of OM (see Supplementary Figure 3). The EET  
378 here obtained is consistent with a value of 83 km previously reported (Taylor et al., 2020) for the  
379 elastic thickness of the lithosphere near Olympus Mons.

380 In order to explain the rise of the break in slope to its actual altitude of about 7 km, we  
381 tested different plume models. The topography resulting from adding the effect of a plume with a  
382 thickness of 650 km and a temperature difference of 300 K to the initial deformation gave the  
383 best result and is shown in Fig. A3. Due to the large width of this plume (radius of 1000 km), its  
384 topographic effect is similar for all tested EET and the maximum topography reaches 15 to 17 km.  
385 Since the buoyancy is the product of thickness and density contrast, a thinner plume with higher  
386 temperature or a thicker one with smaller temperature would give the same results.

387 Figure S4 shows details of the model and resulting crustal deformation due to different  
388 mechanisms (topographic load, underplating, plume) as well as their effect on resulting surface  
389 topography for the model with an EET of 80 km. For topographic load and underplating, the sum  
390 of resulting surface topography and crustal deformation (taken positive downwards) equals the  
391 thickness of the load. For the plume, crustal deformation equals surface topography. The effects  
392 vary linearly with the height of each mechanism allowing thus to estimate topography for other  
393 load values.

394 Our modeling approach was further applied to the case of Big Island (Hawaii) as a test. A  
395 full investigation with different sets of parameters has been already achieved by others (e.g.,  
396 Watts and Zhong, 2000; Zhong and Watts, 2002, 2013) and is beyond the scope of the present  
397 study. We simply considered here a set of realistic parameters: a topographic load of 13-14 km,  
398 an EET of 30 km, crustal underplating of 40 %, crustal density of 2.8 kg/m<sup>3</sup>, asthenosphere  
399 density of 3.2 kg/m<sup>3</sup>, plume excess-temperature of 200°C, plume radius of 250 km, and plume  
400 thickness of 200 km. The predicted surface topography (Fig. S5) is overall consistent with the  
401 observed topography. Downward lithospheric flexure by volcano load produces an extended  
402 basin about 2-3 km deeper than the present topographic surface of the seafloor along the  
403 Hawaiian Deep. However, the latter comprises a >2 km-thick sedimentary sequence that  
404 gradually filled the basin during main island evolution (e.g. Moore et al., 1994, and references  
405 herein). Therefore, modeled basin depth is consistent with the actual depth of the basement  
406 (under the sediments), which confirms that our numeric approach reasonably predicts first-order  
407 basement deformation.

408

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571

572 **Authors contribution**

573 AH conceptualized the study, performed the geomorphological analyses, and wrote the first draft.

574 HZ conducted the numerical modelling of plate deformation. FS and SB took of care of the

575 aspects related to the age and extent of Martian Oceans and formation of the Tharsis Bulge. All

576 authors contributed to scientific discussions/inputs, and drafting of the article.

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579 significantly improve the manuscript. This is LGMT contribution 180.

580

581 **Table captions**

582 Table 1: Parameters used in the calculations of elastic plate deformation

583

584 **Figure captions**

585 Figure 1: Upper inset: Location of Olympus Mons (OM) at the NW edge of the Tharsis Rise  
586 (colour map and NW-SE general topographic profile from global MOLA DEM, vertical  
587 exaggeration by a factor 17). Main Figure: Morphological characteristics of Olympus Mons from  
588 MOLA data. (a) Shaded-relief map (illumination from top), (b) slope map, and (c) topographic  
589 profiles (vertical exaggeration by a factor 9)

590

591 Figure 2: Morphological characteristics of Pico, Fogo, and Big Island. Upper panels: shaded-  
592 relief maps (vertical illumination). Island coasts are drawn with thick black contours. The trace of  
593 the main lateral flank failures is shown with dashed lines. Straight lines: trace of topographic  
594 sections exposed in lower panels. Insets: location of the 3 islands on a cropped World map  
595 (Lokal\_Profil, Public Domain). PV: Pico Volcano; FV: Fogo Volcano; ML: Mauna Loa, MK:  
596 Mauna Kea; K: Kilauea; HFS: Hilina Fault System. Lower panels: topographic cross-sections  
597 (profile traces in upper panels).

598

599 Figure 3: (a) Fossil passage zone from sub-aerial to submarine lava flows on the island of Santa  
600 Maria (Azores). Pl: pillow-lavas. (b) Close up on a pillow breccia in the submarine units at  
601 present shore.

602

603 Figure 4: New model of evolution proposed for Olympus Mons. The inferred ages (in Ga) are  
604 indicative. Vertical scale shows present elevation from MOLA data. The amount of subsidence  
605 and surface uplift is derived from our new modelling of (1) lithosphere flexuration under early  
606 volcanic load in the presence of an ocean (stages a and b) and inferred partial basin infill by

607 detrital sediments (grey colour) and (2) volcano uplift triggered by plume load and/or magma  
608 underplating during emplacement of the Tharsis Rise (stage c) and/or late Amazonian activity  
609 (stage d). See discussion in the main text.

610

611 Figure 5: Main morphological characteristics of Alba Mons from MOLA data. (a) Shaded-relief  
612 map (illumination from W); (b) Topographic profiles from the centre to the outer parts of the  
613 volcano (see profile traces in (a)).

614

Parameter	Value
E	100 GPa
$\nu$	0.25
$T_e$	30, 50, 80, 100 and 150 km
$\rho_c$	3200 kg/m <sup>3</sup>
$\rho_m$	3500 kg/m <sup>3</sup>
$\Delta T$ plume	300 K
$g$	3.71 m/s <sup>2</sup>

Hildenbrand et al., Table 1

Figure 1

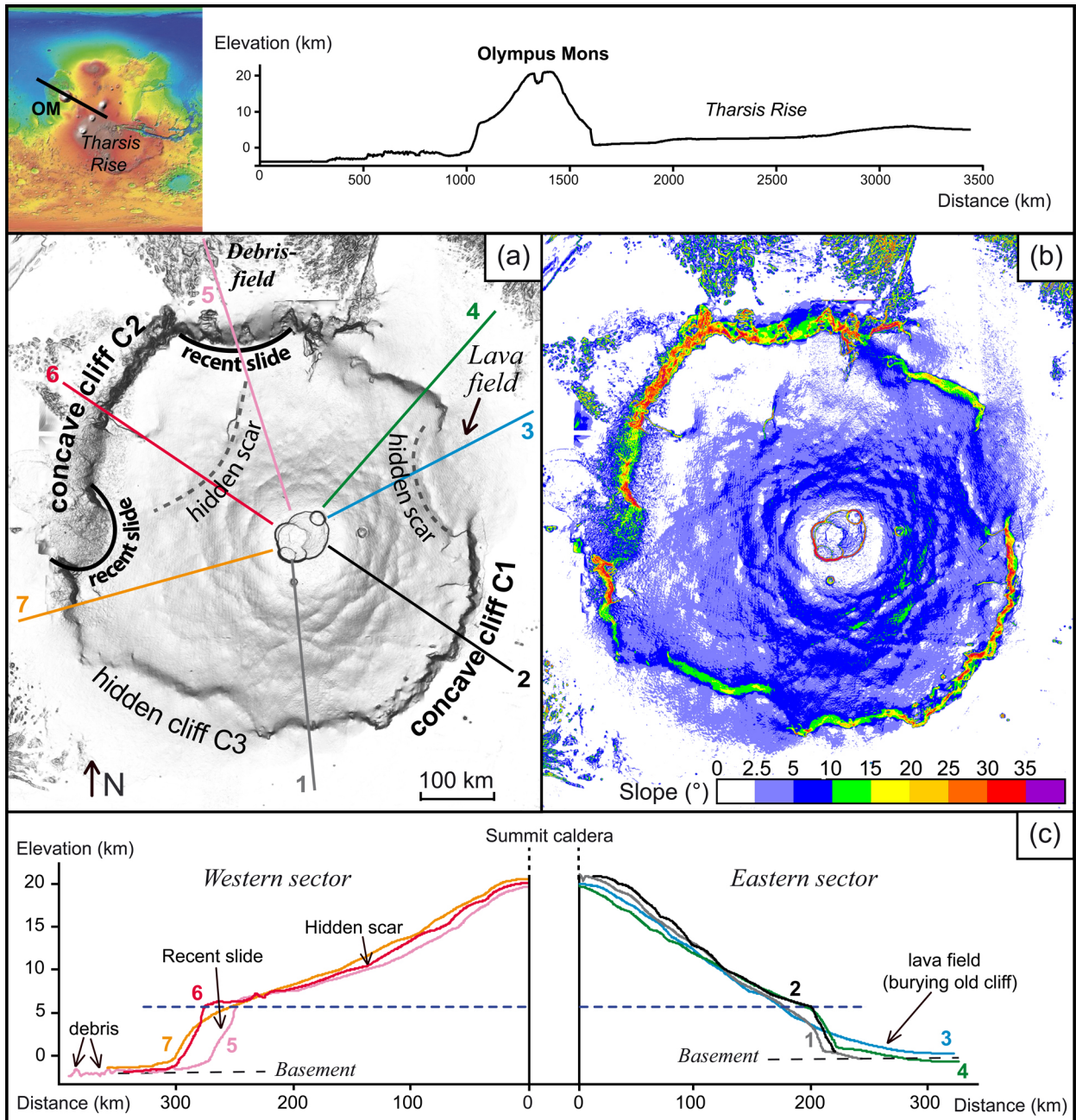


Figure 2

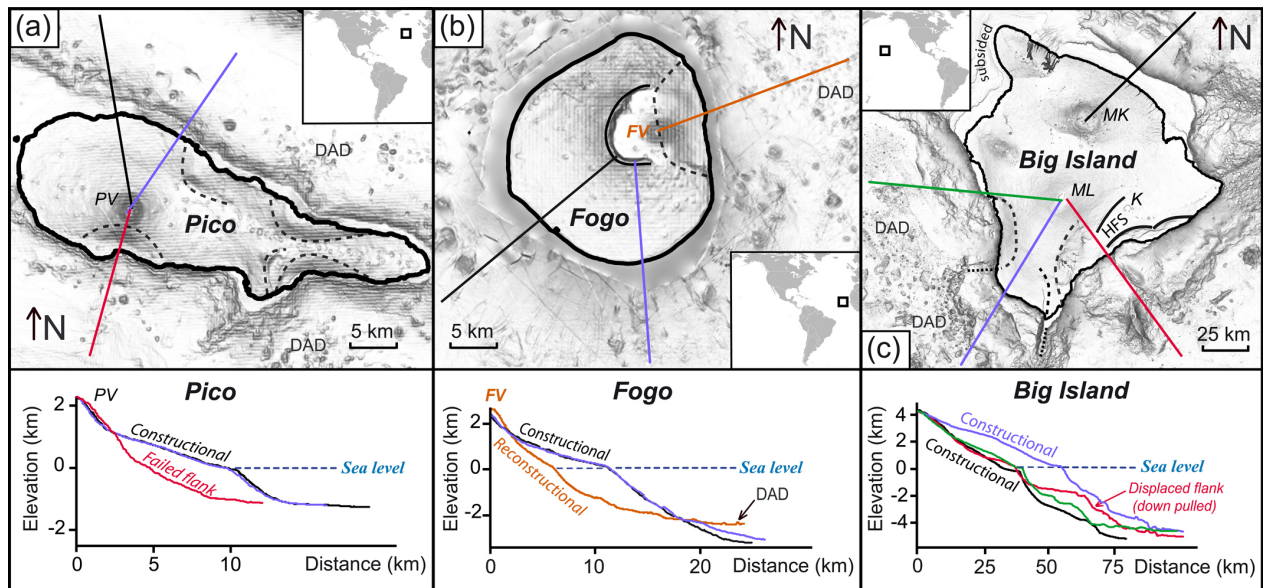


Figure 3

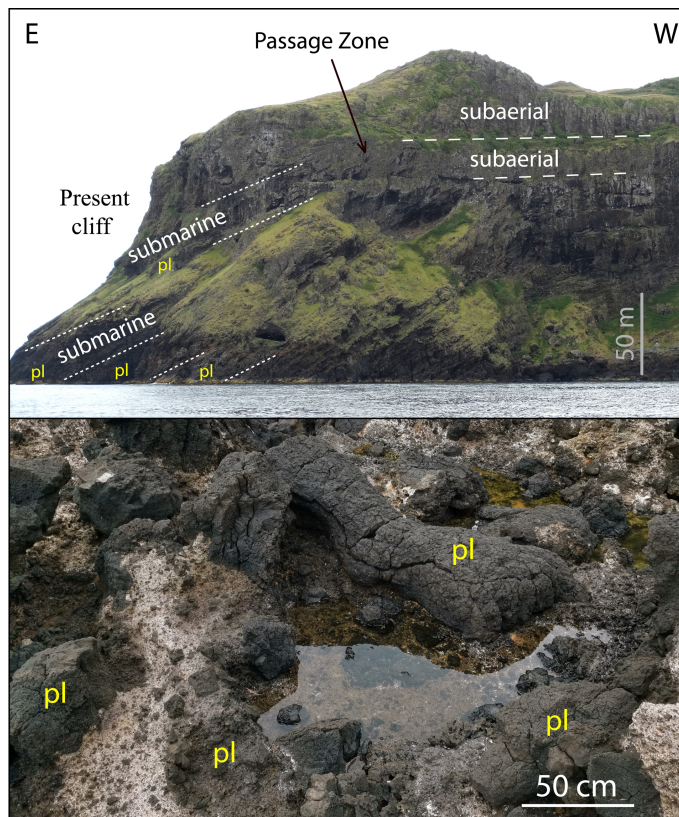




Figure 4

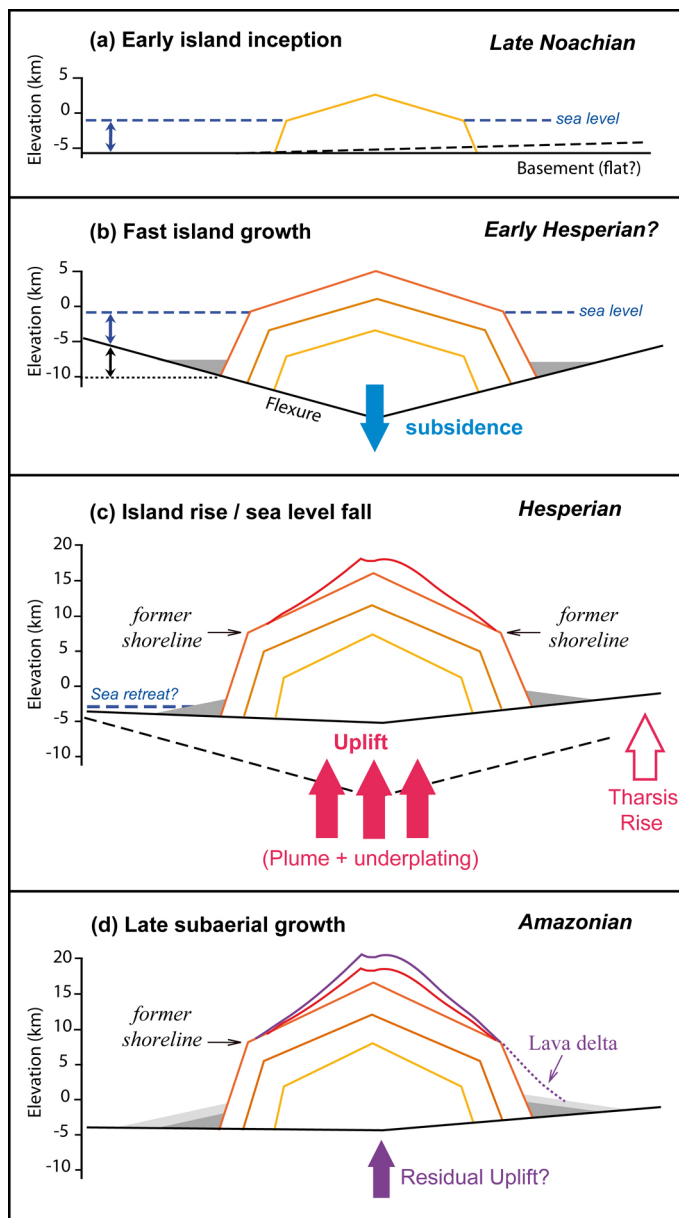


Figure 5

