

Abstract:

Hesperian Mars was characterized by a unique style of geodynamic activity that left crucial volcano-tectonic records in the form of extensive flood lavas covered by wrinkle ridges. Yet, little is known about the context of their formation. Here, we perform a tectonic and geophysical investigation of Hesperia Planum, a 1700-km-diameter volcanic plain covered by wrinkle ridges. Our tectonic analysis reveals that the planum has the highest density of wrinkle ridges on the planet and a characteristic compressional peak strain of about 3.20×10^{-3} , almost 2 times larger than typical Hesperian compressional strains. We invert gravity and topography data and find that simple crustal loading and volcanism cannot explain the tectonic record. An additional source of deformation is thus required. We demonstrate that a loading sequence of plume-induced uplift, volcanism, and subsidence, following an evolutionary path similar to flood basalt provinces on Earth better fits the observations. This plume model is able to explain the peak strain, crustal thinning, and low relief of Hesperia Planum. The inferred plume head size (~ 1400 km) and temperature anomaly (~ 320 K) are consistent with large terrestrial plumes. Based on a fit to the tectonic record, we determine a plume center location that correlates with a cluster of wrinkle ridges, local crustal thinning, and a circular magnetic low, where the latter could be the result of a thermal demagnetization of the lithosphere in the presence of the ascending plume. Our analysis suggests that scattered mantle plumes could be at the origin of Hesperia Planum and other late Noachian to early Hesperian volcanic provinces within the highlands.

44 **1 Introduction**

45 Volcanic products are testimonies of a planet's internal activity and can be used
46 to infer its geologic history (L. Wilson & Head, 1983; Greeley & Schneid, 1991). On
47 Earth, surface manifestations of different aspects of mantle dynamics are found in the
48 form of plate tectonics (T. J. Wilson, 1965) and the presence of plume-induced
49 hotspots (Morgan, 1971) and large igneous provinces (Richards et al., 1989; Coffin &
50 Eldholm, 1994). A large igneous province is commonly thought of as a vast ($>10^5$
51 km²) and low-relief effusive intraplate volcanic deposit (Coffin and Eldholm, 1994;
52 Ernst, 2014; Ernst et al. 2021), which was emplaced over a relatively short period of
53 time (few tens of Myr, Richards et al., 1989). Tens of large igneous provinces and their
54 associated flood basalts have been found all across the Earth and are thought to have
55 formed throughout our planet's history (see Ernst & Buchan, 2001, for a review).

56 The formation of large igneous provinces entails upwelling and decompression
57 melting of mantle material, but the driving force is at the center of lively controversies
58 and debates (see Davies, 2000; Foulger, 2010, for a review). A dynamic scenario
59 suggests that large igneous provinces and flood basalts are due to ascending plumes
60 of anomalously hot material coming from the lower mantle (Richards et al., 1989;
61 Griffiths & Campbell, 1991; Torsvik et al., 2010). Alternative scenarios do not
62 necessarily require mantle plumes, and instead involve, for example, local mantle
63 dynamics with upper-mantle sources intersecting regions of lithospheric stretching
64 and thinning (e.g., Anderson, 2000; Hales et al., 2005).

65 Flood basalt volcanism is common to all terrestrial planets (Head & Coffin,
66 1997). However, the flood volcanism on the Moon responsible for the nearside
67 maria is linked to regions of thin crust and high concentrations of heat producing
68 elements, and thus is not analogous to terrestrial large igneous provinces. Similarly,
69 volcanism on Mercury is strongly concentrated in the enigmatic northern volcanic
70 province. The planet's relatively thin mantle would hinder the formation of large
71 mantle plumes as understood on Earth (Campbell & Griffiths 1990). For those
72 reasons, the formation of the largest volcanic provinces on the Moon and Mercury
73 are not likely to be associated with mantle plumes as traditionally viewed on Earth.

74 Venus harbors several large volcanic provinces with radiating dike structure
75 resembling the large igneous provinces found on Earth (e.g., Head & Coffin, 1997;
76 Ernst & Desnoyer 2004; Hansen 2007; MacLellan et al., 2021). Mars is a basalt-
77 covered world, with about 40% of the planet's surface being covered by lava flows
78 of Hesperian age (3.7–3.4 Ga, Tanaka et al., 2014). However, the large Tharsis
79 volcanic rise and the vast volcanic plains covering the northern lowlands are unlike
80 large igneous provinces on Earth in scale ($\sim 10^7$ versus 10^5 km²) and evolutionary
81 history, and may be related to hemisphere-scale mantle upwellings (e.g., Redmond
82 & King, 2004). In contrast, the southern highlands harbor several discrete late
83 Noachian and early Hesperian volcanic plains covered by wrinkle ridges that are
84 similar in scale to large igneous provinces on Earth. The high eruption rate that
85 characterized this plains-forming volcanism is further reminiscent of that in plume-
86 induced flood basalts and large igneous provinces on Earth (e.g., Head & Coffin,

1997; Mège & Ernst, 2001; Head et al., 2006). Mars has spent most, if not all, of its evolutionary path in a stagnant lid regime without plate tectonics (Solomon, 1978), and with minimal surface erosion for the past 3.5 Gyr. For that reason, the planet holds a more pristine record of its geologic history than the Earth, including that of the formation and evolution of its volcanic provinces. Geologic and geophysical observations on Mars could thus be used to reveal the details of the formation of these provinces, thereby shedding light on the influence of mantle dynamics on magmatism. Although the unprecedented volcano-tectonic activity recorded on the Noachian-Hesperian wrinkle ridged plains may provide the best analogs to large igneous provinces on Earth, the predictions of the mantle plume hypothesis have not yet been explicitly tested against the available geophysical data.

In this study, we perform a tectonic and geophysical investigation of Hesperia Planum, a 1700 km-diameter volcanic plain covered by wrinkle ridges located in the southern hemisphere of Mars. This planum shares similarities to some terrestrial flood basalts and has the highest density of wrinkle ridges on the planet. We start by presenting the general geologic characteristics of the Noachian-Hesperian ridged plains on Mars and how they compare to terrestrial flood basalt provinces. Next, we discuss the topography, gravity, magnetic, and tectonic data at Hesperia Planum. We then present our geophysical approach to determine how the planum formed, including gravity and topography analyses, and a comparison of the tectonic strain

to predictions of a mantle plume model. In the last section, we discuss the implications of our results for the formation of Hesperia Planum.

2 Background and previous work

2.1 Noachian-Hesperian ridged plains

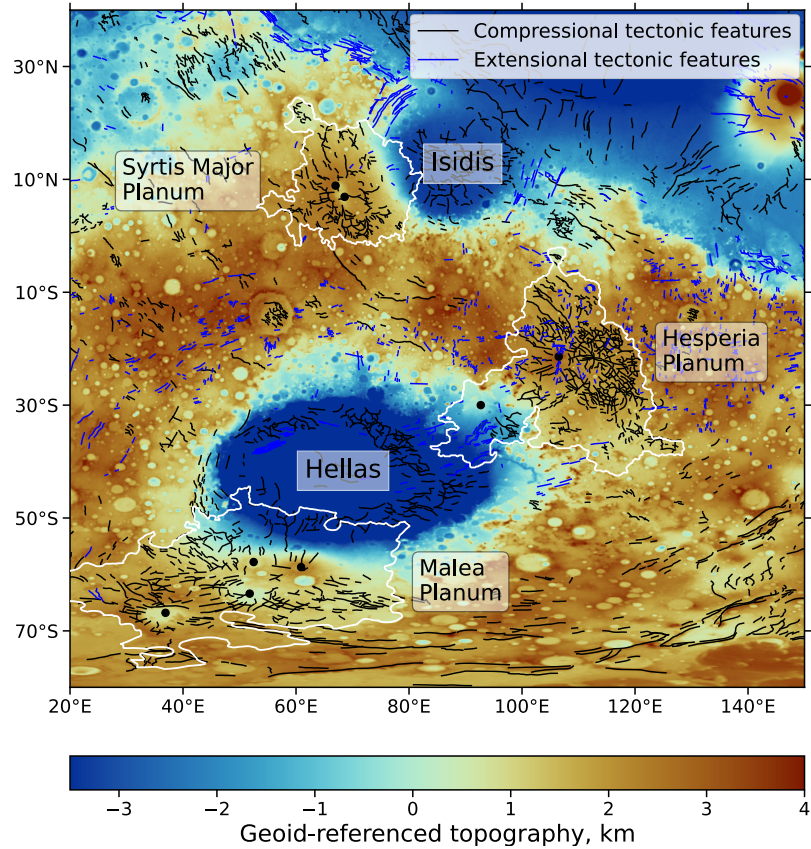
The southern highlands of Mars harbors two large early Hesperian volcanic ridged plains near the Hellas impact basin (Tanaka et al., 2014), namely Syrtis Major (10°N, 60°E) and Hesperia Plana (20°S, 105°E), which have diameters of 1400 to 1700 km (Figure 1). A third major ridged plain, Malea Planum (65°S, 60°E) is located south of Hellas and displays similar overall characteristics as the Hesperian ridged plains, though is slightly older with a Late Noachian age (3.84–3.7 Ga; Tanaka et al., 2014). In the western hemisphere Hesperian lava plains occur within Tharsis, such as Syria and Lunae Plana (Scott & Tanaka, 1986; Golombek & Phillips, 2009). Hesperian ridged plains also cover the northern lowlands, beneath a thin sedimentary layer (Head et al., 2002). However, the Noachian-Hesperian highland volcanic provinces (Hesperia, Syrtis Major, and Malea Plana) formed over relatively short time spans (<200 Myr, Williams et al. 2010; Robbins et al. 2011) with little to no subsequent plains-forming volcanism. This high volcanic pulse makes them distinct from the ridged volcanic plains within Tharsis and covering the northern lowlands that display a more steady volcanic history unique to those particular settings. In this work, we focus on the Noachian-Hesperian highland volcanic provinces (Hesperia, Syrtis Major,

and Malea Plana, with Hesperia Planum taken as the archetype) as potential analogs to continental flood basalt provinces on Earth (Head & Coffin, 1997).

The Noachian-Hesperian highland volcanic provinces sit in subtle 1 to 1.5 km depressions with respect to the surrounding highlands (Figure 1). Such depressions cannot be explained by the flexural response to surface volcanic loading only, as this would not push the volcanic surface below the pre-existing surface. An additional source of crustal modification is thus required to form these ridged plains beyond the observed volcanism alone. These volcanic provinces are further characterized by smooth lava flows with a high density of long (tens to hundreds of km) and sinuous wrinkle ridges (Knapmeyer et al., 2006), which form by folding of near-surface layered volcanic flows above blind thrust faults (Schultz, 2000). The orientation of the ridges as well as the associated strain are related to loading and deformation of the lithosphere and provide important clues on the geodynamic context of their formation (e.g., Mège & Ernst, 2001; Montési & Zuber, 2003; Golombek & Phillips, 2009; Bethell et al., 2021).

The large concentration of wrinkle ridges on the Noachian-Hesperian highland volcanic provinces contrasts with tectonic features observed in the volcanic shields of the Tharsis province. The Tharsis Montes are associated with enormous compressional flexural strain (Banerdt et al., 1982), but little to no compressional tectonics (Golombek & Phillips, 2009). This observation can be explained by a long history in which loading is in equilibrium with flexure, and where the uppermost layers corresponding to the latest eruption phase are close to strain-free. In

150 comparison, the high density of wrinkle ridges on the Noachian-Hesperian highland
 151 volcanic provinces would require a shorter burst of volcanism with flexural
 152 deformation following the eruption, rather than volcanism and flexure operating in
 153 parallel.



154

155 **Figure 1** Surface elevations of the Hellas region. Three major volcanic units are
 156 displayed in white, including Syrtis Major, Hesperia, and Malea Plana (Tanaka et al.,
 157 2014). Two large impact basins, Isidis and Hellas are also referenced. Tensile structures
 158 (blue) and compressive features (black) as mapped by Knapmeyer et al. (2006) are
 159 annotated. Black dots indicate the positions of volcanic edifices, including Nili (67.0°E,
 160 8.9°N), Meroe (68.6°E, 6.9°N), Tyrrhena (106.5°E, -21.4°N), Hadriaca (92.7°E, -30.0°N),
 161 Peneus (52.5°E, -57.8°N), Amphitrites (60.9°E, -58.7°N), Malea (51.8°E, -63.4°N), and
 162 Pityusa (36.9°E, -66.8°N) Montes. Topography data (Smith et al., 2001) are referenced
 163 to the geoid (Genova et al., 2016).

2.2 A comparison with terrestrial flood basalts

The most pristine flood basalt province on Earth is the 800-km diameter, young (~17 Ma), Columbia River Basalt Group in western North America (see Reidel et al., 2013, for a review). The cause of the flood-basalt volcanism has been controversial (e.g., Hales et al., 2005; Hooper et al., 2007), but a plume origin has been found to best explain the geophysical and geochemical characteristics of the province (see Camp, 2013, for a review). The Columbia River Basalt Group shares interesting similarities with the Noachian-Hesperian ridged plains on Mars, including that it also sits in a kilometer-scale broad depression, displays a local crustal thinning (Laske et al., 2013), and is the only place on Earth in which wrinkle ridge-like thrust faulting has been documented (Hooper & Conrey, 1989; Watters, 1989; Mège & Ernst, 2001). These features are testimonies of regional subsidence of the plateau, which is thought to be correlated with the rate of volcanic eruption (Reidel et al., 1989, 2013), and could be linked to the deflation of the plume source. Furthermore, whereas typical plume-induced flood basalts on Earth led to rifting and often transitioned to seafloor spreading and ocean opening (Courtillet et al., 1999), the Columbia River Basalt Group has not reached this final stage and is thus only moderately affected by plate tectonics. One main difference between the Noachian-Hesperian highland volcanic provinces and the Columbia River and other flood basalts on Earth is the overall absence of major dike structures on the former, which are often associated with the effects of mantle plumes (e.g., Ernst & Buchan, 2001). Some local dike systems have

186 been observed at Hadriaca Mons (Korteniemi et al., 2010), and western Terra
187 Tyrrhena (Head et al., 2006), but these are significantly smaller than the large
188 Mackenzie dike swarm in Canada with radiating dikes up to 3000 km long, or the
189 Chief Joseph dike swarm in the Columbia Plateau (Mège & Ernst, 2001; Buchan &
190 Ernst 2021). Radiating dike swarms require a horizontally extensional stress regime,
191 which may not have existed on Hesperian Mars away from the Tharsis volcanic rise.
192 This suggests that if large dike systems existed on the Noachian-Hesperian highland
193 volcanic provinces, they have been confined to the area of the plains and buried
194 beneath voluminous flood basalts.

195 These overall similarities allow us to suggest that the Noachian-Hesperian ridged
196 plains on Mars could have formed following an evolutionary path similar to large
197 igneous provinces on Earth, and we will here focus on the plume hypothesis. The
198 formation and evolution of a mantle plume is controlled by the thermophysical
199 characteristics of the lithosphere and mantle (Campbell & Griffiths 1990). Earth and
200 Mars are different in many respects, and this affects the generation of mantle plumes.
201 The dimensions of the plume head are primarily controlled by the height of rise
202 through the mantle, gravity, and viscosity of the mantle. Using scaling relationships
203 based on laboratory experiments (Campbell and Griffiths, 1990), it is possible to show
204 that for similar viscosities, the diameter of a plume head on Mars would be only about
205 15% smaller than on Earth, since the effects of gravity and rise height (mantle
206 thickness) partially cancel one another out. If instead the martian mantle viscosity
207 were two orders of magnitude different, the diameter would be 0.3 to 2 times larger

on Mars, where higher viscosities lead to larger plume heads (Text S1). This scaling suggests that similar plume head geometries could be found on Earth and Mars.

2.3 An idealized sequence of plume-induced flood basalt volcanism on a one-plate planet

The origin of large igneous provinces on Earth is still debated, with decompression melting being related to either plate tectonic processes or to the ascent of a deep mantle plume (see Davies, 2000; Foulger, 2010, for a review). For example, the North Atlantic Igneous Province is thought to have formed following rifting-induced delamination (Petersen et al., 2018), while the Columbia River Basalt Group may have formed from a deep mantle plume (Camp, 2013). Without plate tectonics on Mars, non-plume alternatives for the generation of volcanism involve passive scenarios such as partial melting below a thickened crust and melting due to small-scale mantle convection (or edge-driven convection), as well as active mechanisms such as eruption enabled by lithospheric extension and delamination. Eruptions driven by extension are incompatible with the dominantly compressional stress regime away from Tharsis during the Hesperian and the absence of regional extension associated with the Noachian-Hesperian ridged plains (Figure 1). All these mechanisms for generating volcanism would also not generate sufficient, if any, surface uplift and subsidence to explain the tectonic record as discussed below.

While plate tectonics may interact with plumes in the formation of large igneous provinces on Earth, the absence of global plate motions does not preclude the formation of plume-induced flood basalts, as widely inferred to be the case on Venus (e.g., Gülcher et al., 2020). For that reason, although significant uncertainties exist in the formation of large igneous provinces and their flood basalts on Earth, we here present an idealized plume-induced evolution sequence on a stagnant lid planet drawing from models and observations on the Earth (Figure 2). 1) A deep thermal anomaly creates a buoyant mantle plume that rises and causes thermal uplift and extension of the surface (e.g., Griffiths & Campbell, 1991; Farnetani & Richards, 1994; Rainbird & Ernst, 2001; Saunders et al., 2007). 2) The plume head impinges the base of the lithosphere and flattens, at which point thermal uplift is at a maximum (Griffiths & Campbell, 1991; Campbell, 2007) and crustal materials are replaced by plume materials causing a local crustal thinning (White & McKenzie, 1989). 3) Extensive flood basalts extrude through the thinned crust on the uplifted surface at large eruption rates ($\sim 1 \text{ km}^3$ per year, Richards et al., 1989), which may be accompanied by giant dike swarms (Ernst & Buchan, 2001), layered intrusions, and magmatic underplating (Phillips et al. 1990; Thybo & Artemieva, 2013; Ernst et al., 2019). For a typical mantle viscosity of about 10^{20} Pa s , which is consistent with expected mantle viscosities on Hesperian Mars (Plesa et al., 2018), most uplift takes place over 5 Myr and flood basalt eruption lasts until at least 5 to 10 Myr after the surface reaches its maximum elevation (Griffiths & Campbell, 1991; Friedrich et al. 2018). 4) The plume head dissipates, leading to a phase of thermal subsidence

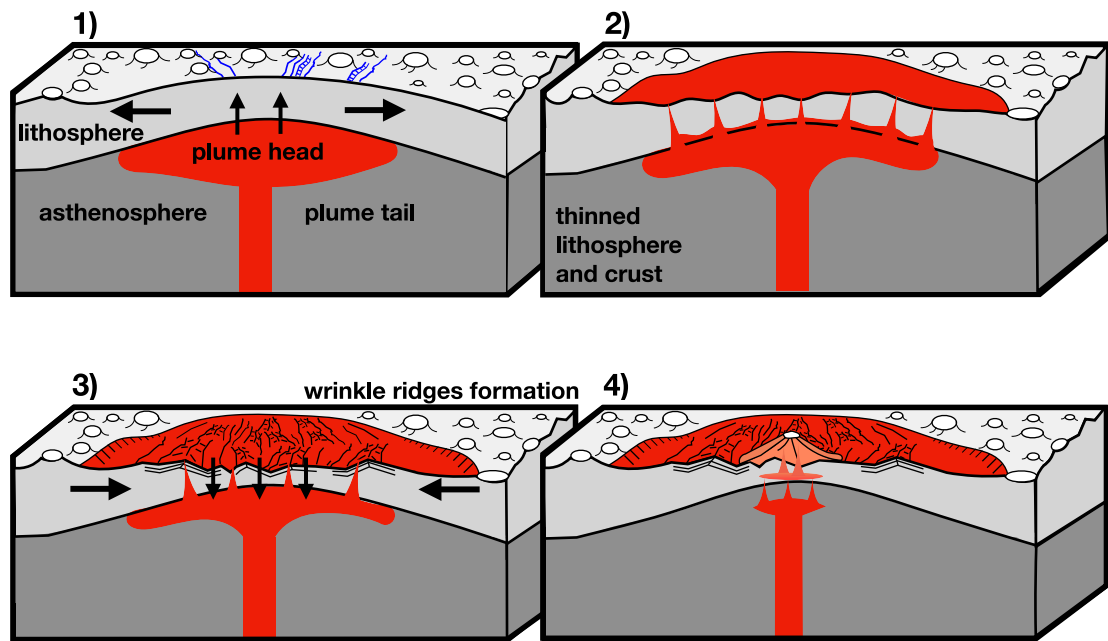


Figure 2 A typical sequence of plume induced flood basalt formation as expected on a one-plate planet, including 1) uplift and extension, 2) flood volcanism, 3) subsidence and compression, and 4) continued volcanism and shield building (see text for details).

250 and lithospheric thickening. With a lack of thermal support, the province transitions
 251 to net downward flexural loading from the prior volcanism and crustal thinning,
 252 leading to a state of compression (Campbell, 2001). 5) Flood basalt volcanism ends,
 253 but plume tail materials may continue to flow upward from the conduit and may
 254 form a volcanic complex and thermally erode the lithosphere (Davies 1994). On the
 255 Earth, plume-induced uplift often triggers rifting and continent breakup following
 256 stage 3 above, and continued plume-tail activity creates a hotspot chain of volcanic
 257 islands in the over-riding plate. In contrast, on a stagnant lid planet, all of the above
 258 stages would be recorded in the same geographic region.

The most important aspect of this idealized sequence for this study is that most of the flood basalts form in a stress-free condition on the uplifted surface, followed by later subsidence. Therefore, the uppermost lava flows will not record tensile but compressional strains as the plume-induced uplift subsides down. On Earth, subsidence would coincide with or follow rifting, such that compression is typically not recorded on the flood basalts (Mège & Ernst, 2001). The plume-induced thermal uplift is reversible, and the strains accumulated during uplift will be proportional to the later compressional strains recorded on the flood basalts. Some anelastic effects will also affect the uplift and later subsidence as expressed in the underlying lithosphere (e.g., crustal thinning and tectonic failure, see section 4.3).

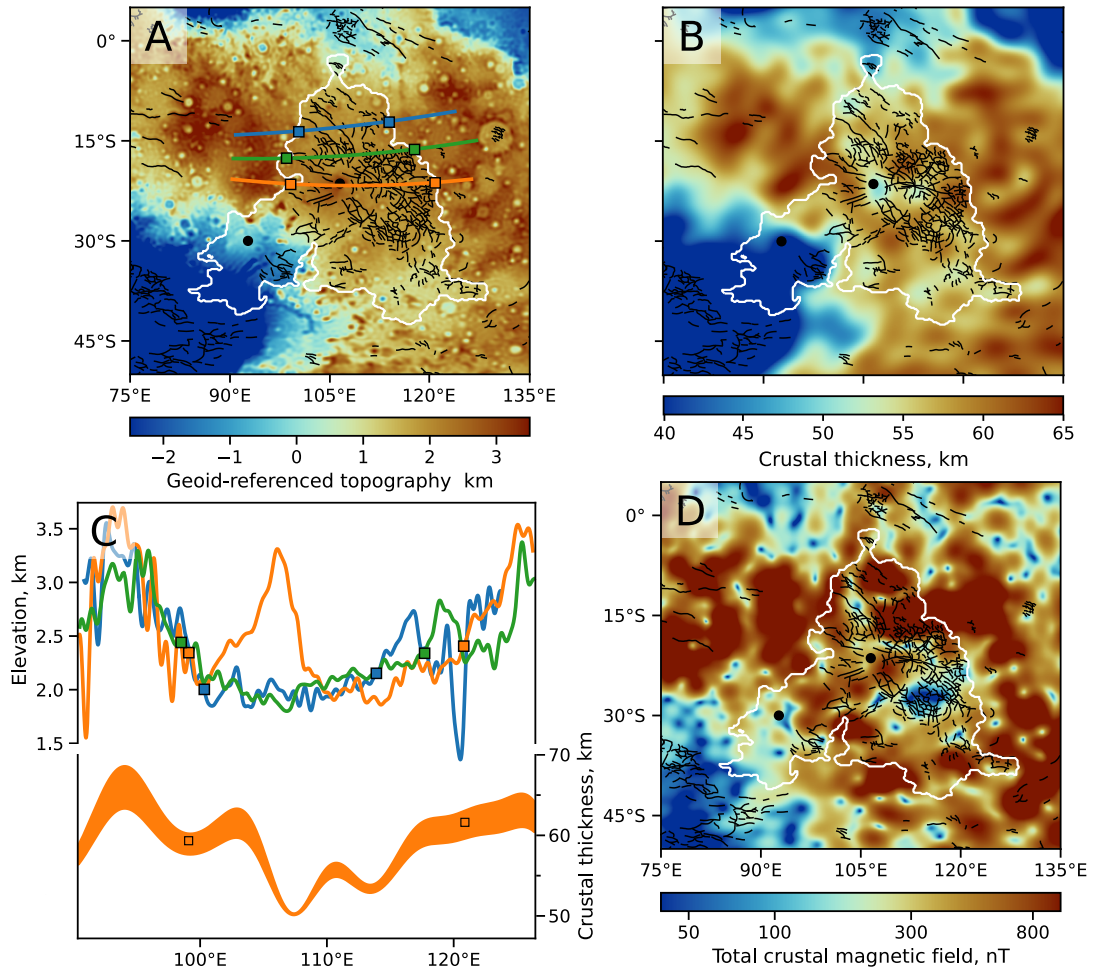
3 Hesperia Planum observations

3.1 Topography, gravity, and magnetic field

Hesperia Planum is a 1700-km diameter volcanic plain that sits in a 1 to 1.5 km depression with respect to the surroundings, and that is superimposed by two low-relief volcanic shields, Tyrrhena (106°E -21°N) and Hadriaca (93°E -30°N) Montes (Figure 3A). A geologic investigation by Ivanov et al. (2005) also noted the depression and observed that valley networks generally flow towards the planum and are buried by the ridged plains. That study proposed that the depression was present before the flood lavas erupted, and is thus not associated with the formation

of Hesperia Planum. Instead, a sequence of massive magmatically induced erosion of the volatile-saturated regolith was proposed.

However, we note that there are a few issues to the scenario proposed in the study of Ivanov et al. (2005). Although valley networks around the periphery do flow toward the planum, it is the ten-kilometer scale roughness of the southern highlands that controls valley orientations, as that creates steeper slopes ($\sim 0.5\%$; Kreslavsky & Head, 1999) than the long-wavelength depression of Hesperia Planum ($\sim 0.1\%$). As such, it is difficult to interpret long-wavelength paleotopography based on river valley networks in the southern highlands. No reorientation of fluvial valleys is expected for any plume uplift/subsidence sequence. Locally, the edge of the volcanic flows will tend to abut against local topographic highs and embay any valley networks, and this relationship is expected even if Hesperia Planum were uplifted during volcanism. Moreover, erosion of 1 km by melting of ground ice would require a regional massive ice deposit of 1 km in thickness, or melting of 10% pore ice down to a depth of 10 km. Deposition of this material within Hellas would then require transport by ~ 1500 km across a region with no net slope toward the basin. The outflow channels themselves do not show evidence for affecting an area beyond the Hellas rim.



298

299 **Figure 3** Surface elevation, crustal thickness, and magnetic field at Hesperia Planum.
 300 (A) Surface elevations with color-coded profiles (Smith et al., 2001). (B) Representative
 301 crustal thickness model assuming a crust density of 2800 kg m^{-3} and mean thickness of
 302 50 km. (C) Extracted elevation profiles across Hesperia Planum, left axis, and example
 303 crustal thickness models for the profile shown in orange, right axis. The crustal
 304 thickness models are from Wieczorek et al. (2022) and assume crustal densities of 2700
 305 to 2900 kg m^{-3} . (D) Total crustal magnetic field with a logarithmic color scale (Langlais
 306 et al., 2019). For geologic context, the images are annotated with a white contour of
 307 Hesperia Planum, black lines for compressive tectonic features, and black dots for
 308 Tyrrhena (106°E - 21°N) and Hadriaca (93°E - 30°N) Montes. Squares on the profiles
 309 denote the edges of the planum. Most of the compressive tectonic features are wrinkle
 310 ridges.

The gravity field at Hesperia Planum is also inconsistent with surface erosion to form its low-relief (Figure S1). If the depression were a result of erosion, the planum should be characterized by a long-wavelength gravity low with respect to the surrounding highlands (~ 120 mGal lower, assuming 1 km of erosion of materials with a density of 2800 kg m^{-3} , see also Evans et al., 2010). Instead, Hesperia Planum's mean gravity field is similar to that of the surrounding highlands both having values of 60 ± 50 mGal (Figure S1). Crustal thickness models that satisfy gravity, topography, and InSight measurements (Wieczorek et al., 2022) further predict a long-wavelength crustal thinning within Hesperia Planum, with a crust that is on average 3 to 5 km thinner than in the surrounding highlands (Figure 3B and C). Alternatively, the gravity signature of the planum could be interpreted as due to a local increase in crustal density ($\sim 100 \text{ kg m}^{-3}$) on a flat crust. The loss of pore space associated with the plume-induced heating could have contributed to increasing the local bulk density of the crust. This implies a weak lithospheric support of local bottom loading to compensate for the low-relief, rather than surface erosion. We conclude that it is thus unlikely that the bulk of the planum's depression is due to surface erosion, though some erosion has occurred at the rim of Hellas (Tanaka et al., 2002).

At Hesperia Planum, crustal thinning reaches its maximum value of up to 10 km underneath Tyrrhena Mons (Figure 3B and C), which may be the result of using a crustal density lower than expected locally at Tyrrhena Mons (about 3200 kg m^{-3} , Broquet & Wieczorek 2019), but could also include contributions from a

magma chamber below the volcano. A second focused zone of crustal thinning east of Tyrrhena Mons is near the center of a cluster of wrinkle ridges. There is no topography associated with this zone of predicted crustal thinning, and thus the true cause must lie in the subsurface (e.g., a solidified magma chamber, a zone of reduced porosity perhaps associated with high heat flow, or true crustal thinning). We finally note that magmatic underplating of the crust by low-density mantle materials is thought to have occurred in some plume-induced flood basalts on Earth (Thybo & Artemieva, 2013) and also on Mars (Phillips et al., 1990), and this would induce an apparent crustal thickening. Given that this is opposite to what is observed at Hesperia Planum, underplating is either overcompensated or these materials have been advected back into the mantle and have not lasted to present-day. Volcanism may also leave behind a dense mantle residuum, but this would also likely be advected back into the mantle on short geologic timescales.

Within Hesperia Planum, the magnetic field is seen to vary significantly from a minimum of 10 to a maximum of 2500 nT on the ridged plains, with a mean value of 560 nT, and decreases in the vicinity of Hellas (Figure 3D). Interestingly, a ~400 km circular magnetic low (<50 nT), one order of magnitude lower than the planum's mean magnetic field, located where wrinkle ridges cluster (~114°E 26°S), stands out from the planum and is also surrounded by a larger circular region of somewhat reduced magnetic field strength (<300 nT). This demagnetized zone is consistent with a scenario in which the crust was heated by local intrusions

above the Curie temperature of its magnetic carriers, as is further discussed in section 6.2.

3.2 Fault density and strain

The tectonic record may provide the most relevant record of the geodynamic evolution of Hesperia Planum. The planum is covered by wrinkle ridges, with high concentrations near 114°E 23°S and 116°E 28°S (Knapmeyer et al., 2006, Figure 4). Over Hesperia Planum, the wrinkle ridge density has a mean value of $0.85 \times 10^{-5} \text{ m}^{-1}$, but reaches higher values of $2.73 \times 10^{-5} \text{ m}^{-1}$ (Figure 4A), where the density was obtained by dividing the ridge length by the enclosing area and smoothed using a moving average 240 km-diameter Gaussian window. Based on a similar tectonic analysis for the entire Martian surface, we found that Hesperia Planum is characterized by the highest density of wrinkle ridges on the planet.

The average tectonic strain over the planum is about 1.91×10^{-3} , reaching a peak value of 4.75×10^{-3} (Figure 4B), where the strain was computed as the wrinkle ridges density multiplied by the ridge height and the cotangent of the fault dip, here assumed to be constant with a value of 30° (see Andrews-Hanna, 2020). For this study, we are primarily interested in the wrinkle ridge cluster described above, and we have thus defined a circular 850-km-diameter area that includes the two peaks, where the mean strain is 3.20×10^{-3} (Figure 4B). For comparison, over Terra Cimmeria's late Noachian units, which are of likely volcanic origin and similar in age to Hesperia Planum at the Noachian-Hesperian boundary (Tanaka et al., 2014), we obtained a mean strain of

~ 1.76×10^{-3} . This value is slightly higher than our estimated global average for all Hesperian units, $\sim 1.47 \times 10^{-3}$. Planetary cooling models predict similar contractional strains to those observed outside of Hesperia Planum, on the order of 2.0×10^{-3} since the Noachian-Hesperian boundary, though they depend on modelling assumptions (e.g., Hauck & Phillips, 2002; Andrews-Hanna et al., 2008). Although this range of strain values is consistent with the mean strain at Hesperia Planum, it is only half of that over the characteristic peak strain area. These values reveal that global contraction is not the only process that generated tectonics on Hesperia Planum.

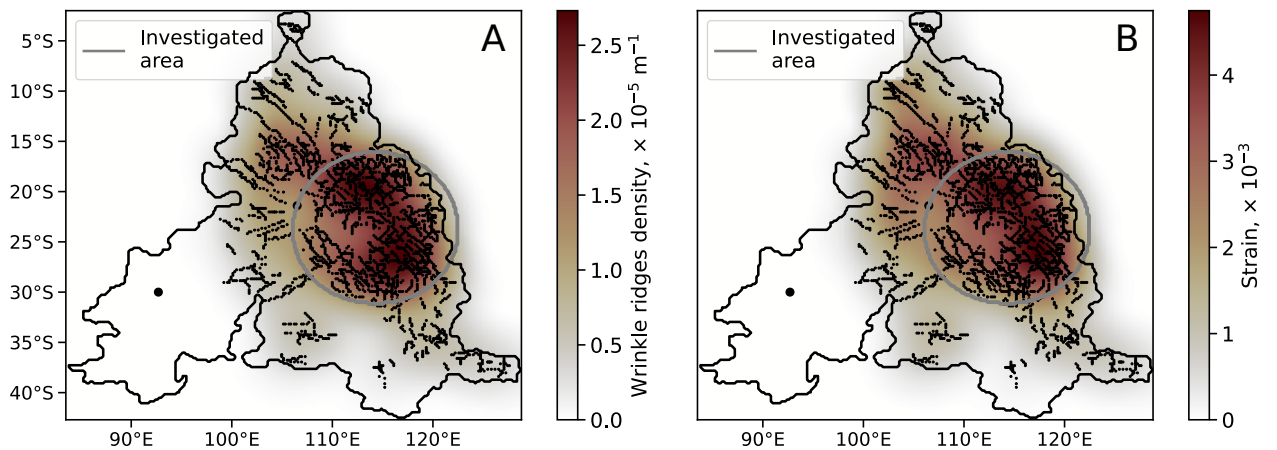


Figure 4 The compressional tectonic record at Hesperia Planum based on the mapping of Knapmeyer et al. (2006), showing the calculated wrinkle ridge density (A) and strain (B). We note that some wrinkle ridges in the vicinity of the Hellas impact basin (near 32°S, 98°E) have been ignored. Black dots indicate the location of Tyrrhena (106°E - 21°N) and Hadriaca (93°E - 30°N) Montes.

3.3 Age relationships

Age relationships between the different structures within Hesperia Planum can be useful to unveil relative formation sequences. The highlands surrounding the planum are thought to be early to middle Noachian (4.1–3.8 Ga) in age and among the oldest surfaces on Mars (Tanaka et al., 2014). The ridged plains of Hesperia Planum have ages ranging from about 3.5 to 3.7 Ga (Williams et al., 2010), whereas age estimates for Tyrrhena and Hadriaca Montes vary from ~3.2 and 4.0 Ga (Williams et al., 2008; 2009; Robbins et al., 2011). At face value, and given the overlapping ages, it is unclear whether these volcanoes formed prior to or concomitant with the volcanic plains and could have acted as the major source for their formation.

However, we note that the youngest ages for Tyrrhena Mons (~3.2 Ga) are likely associated with erosion of the pyroclastic deposits on the flanks rather than the formation of the volcano itself. The oldest ages (3.8 or 4.0 Ga, with the younger age for the stratigraphically lower unit; Williams et al., 2008) rely on a very small number of large impact craters and are thus highly uncertain (4 craters with diameter >4 km in each of two units). The older ages of Tyrrhena Mons and Hesperia Planum overlap at the 1-sigma uncertainty level, and thus should be considered indistinguishable (Williams et al., 2008; 2010). The volcanic shields and the ridged plains of Hesperia Planum further have different age biases. Whereas the high standing relief of a volcano will continue to be exposed to cratering and be less susceptible to volcanic resurfacing, the low-lying flood basalts and ridged plains will always be overprinted by younger

flows. One important observation, however, is that the caldera of Tyrrhena Mons is younger than the lava plains, with a surface age of 3.4 Ga (Robbins et al., 2011). This reveals that a major episode of Tyrrhena Mons' volcanic activity postdates the formation of the volcanic plains of Hesperia Planum, as would be expected in a plume-induced flood basalt sequence (Figure 2).

4 Methods

In this study, we apply three distinctly different geophysical models to investigate the formation of Hesperia Planum. In the first analysis (section 4.1), we probe the structure of the lithosphere by making simplifying assumptions regarding the nature of surface and subsurface loading, varying the governing parameters in an attempt to fit the spectral ratio of observed gravity and topography data (i.e., the admittance, e.g., Broquet & Wieczorek, 2019). This model evaluates whether the regional gravity and topography are consistent with an erosional sequence to form Hesperia Planum, as suggested in Ivanov et al. (2005), or instead require a combination of top and bottom loading as might arise from plume-induced crustal thinning. In the second analysis (section 4.2), we invert observed gravity and topography data for the loading, displacement and associated strain using an updated version of thin-shell model presented in Banerdt et al. (1986). This inversion compares the observed tectonic record to modeled strains to test whether Hesperia Planum can be explained by crustal loading preserved in the gravity and topography data, as might arise for an origin through passive volcanism, such as in

response to rise of shallow mantle material beneath a zone of already thinned crust. The third analysis (section 4.3) explores a dynamic scenario in which a sequence of plume induced uplift and subsidence is imprinted in the tectonic record (see Figure 2). This model represents a mantle plume loading the base of the lithosphere and deforming the surface, together with subsequent volcanism. The comparison of the modeled strains to the tectonic record is described in section 4.4.

4.1 Modelling the gravitational signature of Hesperia Planum

We model the lithospheric response of Hesperia Planum using a localized admittance technique (see Broquet & Wieczorek, 2019 for a full description of the model). The model makes use of observed topography data to predict the surface load, flexure, and associated gravity field given input lithospheric and crustal parameters. Internal loads, such as from crustal thickness variations, are also modelled using a subsurface to surface load ratio. The resultant theoretical gravity field at Hesperia Planum is extracted using a nominal spherical cap localization window centered at 108°E 18°S, and with a diameter and bandwidth of 24° and 10 degree, respectively, for an optimal concentration factor (>99%, see Wieczorek & Simons, 2005). This window was designed to include the bulk of Hesperia Planum, while avoiding the Hellas and Isidis impact basins that experienced distinctly different evolutionary paths.

Given that we are interested in the long-wavelength signature of the planum, our inversion fits the observed admittance and correlation functions, using an RMS

(root mean square) misfit equation, between spherical harmonic degrees 13 to 34, which are equivalent to wavelengths of about 1500 and 600 km. Degrees 10 to 12 have been ignored as their admittance signal is large and likely aliased by long-wavelength loads on the planet such as Tharsis and the planetary flattening (see also Belleguic et al., 2005). We perform an exhaustive grid search to determine the best-fit elastic thickness T_e , crustal density ρ_c , ratio between the surface and subsurface loads f , and phase relationship between these loads α . The load ratio can have values less than, equal to, or higher than 1, which correspond to a top-load dominated case, an isostatic state (bottom load equal surface load), or a bottom-load dominated case, respectively (Broquet & Wieczorek, 2019). In the case of surface erosion with no post isostatic adjustment, the load ratio is expected to be zero (no bottom load). Constant parameters include the mantle density, Young's modulus, and Poisson's ratio, which were set to values of 3500 kg m⁻³, 100 GPa, and 0.25, respectively, following previous admittance studies (Belleguic et al., 2005; Broquet & Wieczorek, 2019).

4.2 Inverting for loading, displacement, and strain from gravity and topography

Whereas the previous section presented a forward model of the gravity field, we here test whether Hesperia Planum, and its characteristic strain record, could have formed from passive volcanism using an inverse approach (Text S2). In this scenario, the geodynamic history of Hesperia Planum, including all loads (mantle or crustal) and associated deformation, is assumed to be fully recorded in the present-day

gravity and topography (unlike the mantle plume model; see below). We invert observed gravity and topography data for the loading, displacement and associated strain using an updated version of thin-shell model presented in Banerdt et al. (1986), which is described in Text S2. This model is similar to classical crust-mantle relief inversions from gravity and topography (e.g., Wieczorek et al., 2022), with the addition that it partitions the crustal thickness variations into top and bottom loads and constrains lithospheric displacement as a function of elastic thickness (Banerdt et al., 1986; Broquet, 2022). The inferred displacement allows us to constrain crustal strains and the direction of faulting, which are compared to the tectonic record. For this model, we assume that there are no internal density variations, which is a common assumption in gravity field inversions (e.g., Wieczorek et al., 2022).

4.3 Modelling a plume induced flood basalt

In a third approach, we explore a dynamic scenario in which flood volcanism at Hesperia Planum originated from a mantle plume source. This scenario assumes that strain during the initial uplift is not imprinted in the tectonic record as it is buried beneath the lava flows (Figure 2). The gravity and topography during uplift are also not preserved, as they would mostly return to their original values after dissipation of the thermal anomaly. However, the compressional strain during subsidence would be recorded in form of wrinkle ridges on the surface of volcanic units, with this strain simply being opposite in sign and equal in orientation and magnitude to the strain during uplift. Any thermal or volcanic erosion of the crust would contribute

to additional subsidence beyond that from the cooling of the lithosphere after dissipation of the plume head. Intrusive magmatism would also induce loading, but this will not be accounted for here, as intrusive loading would directly tradeoff with lava flow loading, and both have unknown thicknesses. Thermal erosion and lateral redistribution of magma away from the plume center would enhance the subsidence of the volcanic plains, as observed in some terrestrial large igneous provinces (Campbell et al., 2001; Ernst et al. 2014). Some magmatically-induced surface erosion could explain the heavily degraded state of the rim of Hellas to the southwest of Hesperia Planum (Tanaka et al., 2002; Ivanov et al., 2005). Thus, the strain predicted by elastic subsidence following cooling of the plume head and overlying lithosphere places a lower bound on the actual subsidence.

This loading model also uses the thin-shell formalism developed in Banerdt et al. (1986; see text S2), modified to work in a forward sense with unknown gravity and topography, but with a known mantle plume head load. The modeled displacement and strains will also be compared to the tectonic record. To represent a plume-induced uplift of the surface, we assume that the plume head load is compensated by dynamic topography only (i.e., no crustal thickness variation), where upward flexure equals upward surface topography. The plume head is modeled as a density contrast, located underneath the elastic lithosphere, and is tapered by a Tukey-like (a cosine tapered step function) window to a given radius. Based on a series of lab experiments simulating mantle plumes (e.g., Griffiths and Campbell 1991; Campbell, 2007), we assume a flattened plume head diameter with

size of 0.5 to 1 times that of the flood basalt province and an aspect ratio in the range of 0.05 to 0.25, yielding a plume head radius of 425 to 850 km and thickness of 85 to 425 km. Estimates on the thermal anomaly associated with terrestrial mantle plume heads range from 100 to 600 K (e.g., White & McKenzie, 1995), equivalent to a density contrast of about 10 to 70 kg m⁻³, assuming a thermal expansivity of $3.3 \times 10^{-5} \text{ K}^{-1}$ (Fei, 1995; Plesa et al., 2018). Following these studies, our inversions explore a wide range of plume head and lithosphere parameters as shown below.

Our dynamic plume model does not account for lithosphere thickness variations during the plume-induced flood basalt sequence. To test the influence of this simplification, we compared the modeled instantaneous flexure at a fixed lithosphere thickness of 80 km to an iterative model with incremental cooling and lithosphere thickening in parallel with the subsidence. The incremental model starts with an initial 350 K plume thermal anomaly underneath a 50-km-thick elastic lithosphere. The thermal anomaly is dissipated linearly, and the elastic thickness is increased accordingly (up to 80 km). The difference between our instantaneous flexure on the 80 km-thickened lithosphere and the incremental model was found to be less than 5% (incremental flexure is slightly higher). However, comparing the instantaneous flexure on the 50 km-thick initial lithosphere and the incremental model gave rise to differences larger than 80% (instantaneous flexure on the thin lithosphere is largely higher). This is simply the result of lithospheric rigidity increasing with the elastic thickness cubed. This result suggests that the best-fit modeled lithosphere thickness will be representative of the thickened elastic

lithosphere at the end of subsidence (or, similarly, the pre-plume background value), rather than the thermally thinned lithosphere during uplift.

Finally, in this theoretical framework, the flood basalt that erupts on the uplifted surface acts as a load and generates additional flexure and compressive strains. Loading, displacement, and strain can be assumed to be linear (Beuthe, 2008), and we thus run a separate model to estimate the deformation contribution coming from the flood lavas. Because there is no topographic signal of central flexure within Hesperia Planum's depression, this independent model assumes that the flexure has been filled by lava flows. We constructed an iterative model where the initial lava thickness is constant everywhere and incrementally increased by flexure until the final surface is flat, as observed today. This model uses as input the initial lava flow thickness, and it is assumed that crustal thickness and internal density variations are negligible. As an example, for an elastic thickness of 100 km, an initial 500 m of lava flows at this scale gives rise to a maximum central flexure of about 1250 m for a maximum final lava thickness of 1750 m. As the thickness of the plains is not well constrained, the initial thickness is varied from 0 to 1 km, along with the elastic thickness of the lithosphere.

4.4 Strain comparison of models to observations

Three independent outputs of our passive (section 4.2) and dynamic (4.3) loading models will be compared to the observed tectonic record. The strain magnitude is compared to observations using the RMS misfit over the restricted

area shown in Figure 4, which includes the two characteristic peak strain concentrations. In a second comparison, the observed and predicted strains are normalized relative to their peak values and compared using the RMS misfit over all of Hesperia Planum. Given that some of the observed wrinkle ridges are in locally orthogonal networks (Figure 4), our models compute the areal strain, which is obtained by adding the sum of principal horizontal strains and their product.

In a third comparison, the principal directions of faulting are compared using the RMS misfit over the whole Hesperia Planum region. Given that wrinkle ridges are generally linear, the fault strike is calculated as the azimuth between the two-end points and the midpoint mapped by Knapmeyer et al. (2006). The predicted principal direction of faulting is obtained by computing the principal strain direction from the modeled horizontal strains. Given the orthogonal tectonic network, both horizontal stresses are expected to be compressional and ridges of either orientation can be in agreement with the model results. Thus, our misfit to the orientation of faulting is measured relative to the closest of the two principal strain directions, with a maximum value of 45° .

Separating these three quantities allows us to focus on specific parameters. For example, the strain magnitude is controlled by the load, which depends on the observed gravity and topography, or the plume head and lava flow thickness, and the density contrast of the plume head. On the other hand, both the normalized strain concentration and direction of faulting will be mostly affected by the elastic thickness of the lithosphere and plume head radius. Finally, we have seen in the

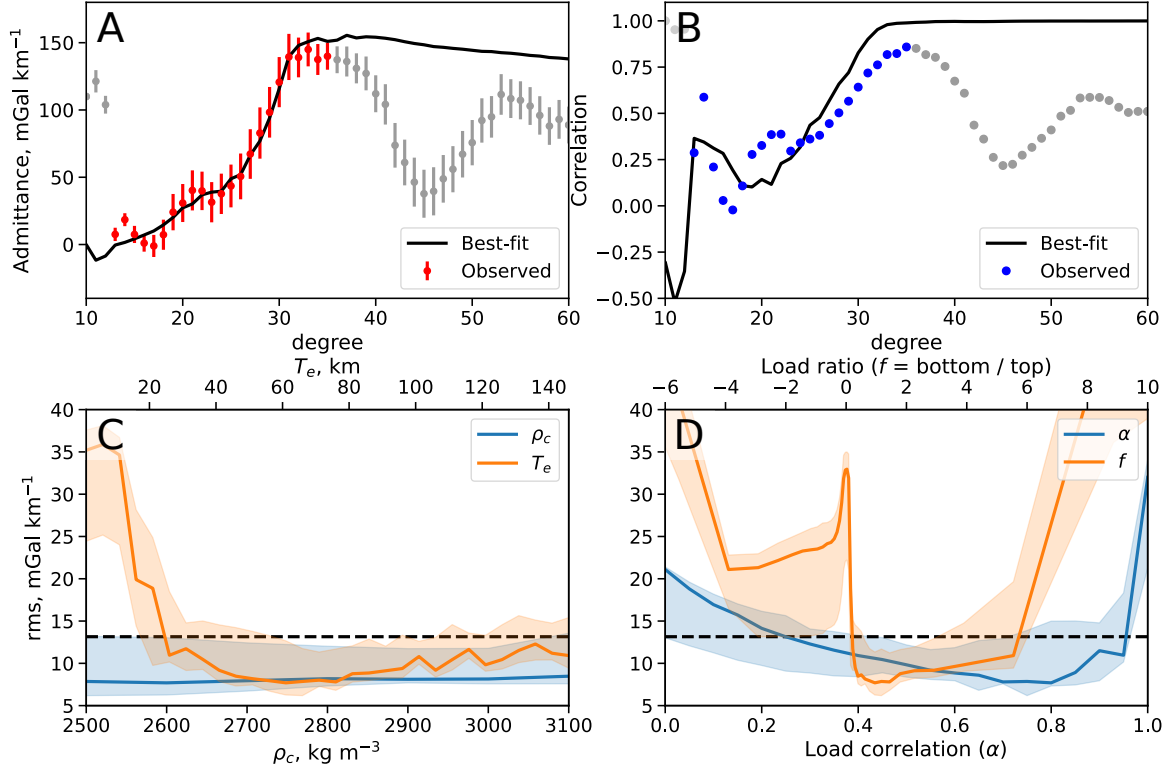
previous section that the peak strain recorded in Hesperia Planum is not located at its center. For that reason, we performed an exhaustive grid search where we systematically varied the position of the plume head for the dynamic model (section 4.3) along with the plume and lithosphere parameters.

5 Results

5.1 The lithospheric properties of Hesperia Planum

We first consider results of the admittance and correlation analysis of the top and bottom loading models, which tests whether the gravity and topography at Hesperia Planum are consistent with an erosional sequence (Figure 5). The best-fit model has load ratio of about 1 and is thus nearly isostatic, though values as low as 0.1 and up to 7 are acceptable. This result indicates that the depression containing Hesperia Planum is dominantly a result of bottom loading (crustal thinning or density increase) rather than erosion as suggested in Ivanov et al. (2005). Formation of the planum's depression by erosion alone would indicate a load ratio of 0 (no bottom load), which corresponds to a sharp peak in the misfit and can be excluded. The elastic thickness is found to be at least 16 km, with a best-fit of about 62 km. The elastic thickness is potentially thicker than locally at Tyrrhena Mons (10 ± 10 km, see Broquet & Wiczorek, 2019), which suggests that the volcano may have formed on a locally thinned lithosphere relative to the thicker lithosphere supporting Hesperia Planum itself. The phase relationship between the surface and bottom load is 0.8 (+0.2, -0.8) and the crustal density is entirely unconstrained, but

the partial decorrelation and lack of crustal density constraints are both expected given the smooth nature of the local topography.



607

Figure 5 Localized admittance and correlation analysis. Fitted observed (A) admittance and associated uncertainty (red) and (B) correlation (blue) with our best-fitting models (black). Data points outside the degree range fit by the model are shaded gray in both plots. (C) Minimum RMS misfit as a function of crustal density (ρ_c , bottom axis) and elastic thickness (T_e , top axis). (D) Minimum RMS misfit as a function of the phase relationship between the surface and bottom load (α , bottom axis) and load ratio (f , top axis). Solid lines give the misfit as a function of one parameter (x-axis) along the minimum misfit for all parameters. The horizontal dashed line in the lower plots indicates the mean uncertainty in the observed admittance from degree 13 to 34. The shaded contour gives the minimum RMS misfit corresponding to variations of the window size ($\pm 2^\circ$), position ($\pm 4^\circ$ in latitude and longitude), and mean crustal thickness of the model (± 10 km).

620

621

5.2 Displacement, and strain from gravity and topography

We next examine the predictions of our passive scenario, in which all loading that has affected Hesperia Planum is reflected in the present-day gravity and topography (excluding the possibility of a past mantle plume). These inversions predict crustal thickness, displacement, and strain, and we note that our crustal thickness models are consistent with Wieczorek et al. (2022). The predicted displacement and strains are dominated by Tyrrhena and Amphitrites Montes (Figure 6), with peak strains of about 1.5 and 6×10^{-3} for elastic thicknesses of 150 and 50 km, respectively. These loading strains are significantly higher than observed at these locations in the tectonic record ($\sim 1 \times 10^{-3}$, Figure 4), which suggest that the two shield volcanoes likely formed over long timespan where flexure operated in parallel to volcanic loading. For an elastic thickness of 50 km (Figure 6A and B), the weak extensional and compressional strains predicted at the observed local cluster of wrinkle ridges in eastern Hesperia Planum are inconsistent with the strong compressional strain observed there. For the larger 150 km elastic thickness (C and D), most of the planum is in compression, but the magnitude of the predicted strain ($\sim 0.45 \times 10^{-3}$) is 10 times lower than observed in the tectonic record at the local strain concentration ($\sim 4.75 \times 10^{-3}$), and this difference cannot be explained by global contraction alone.

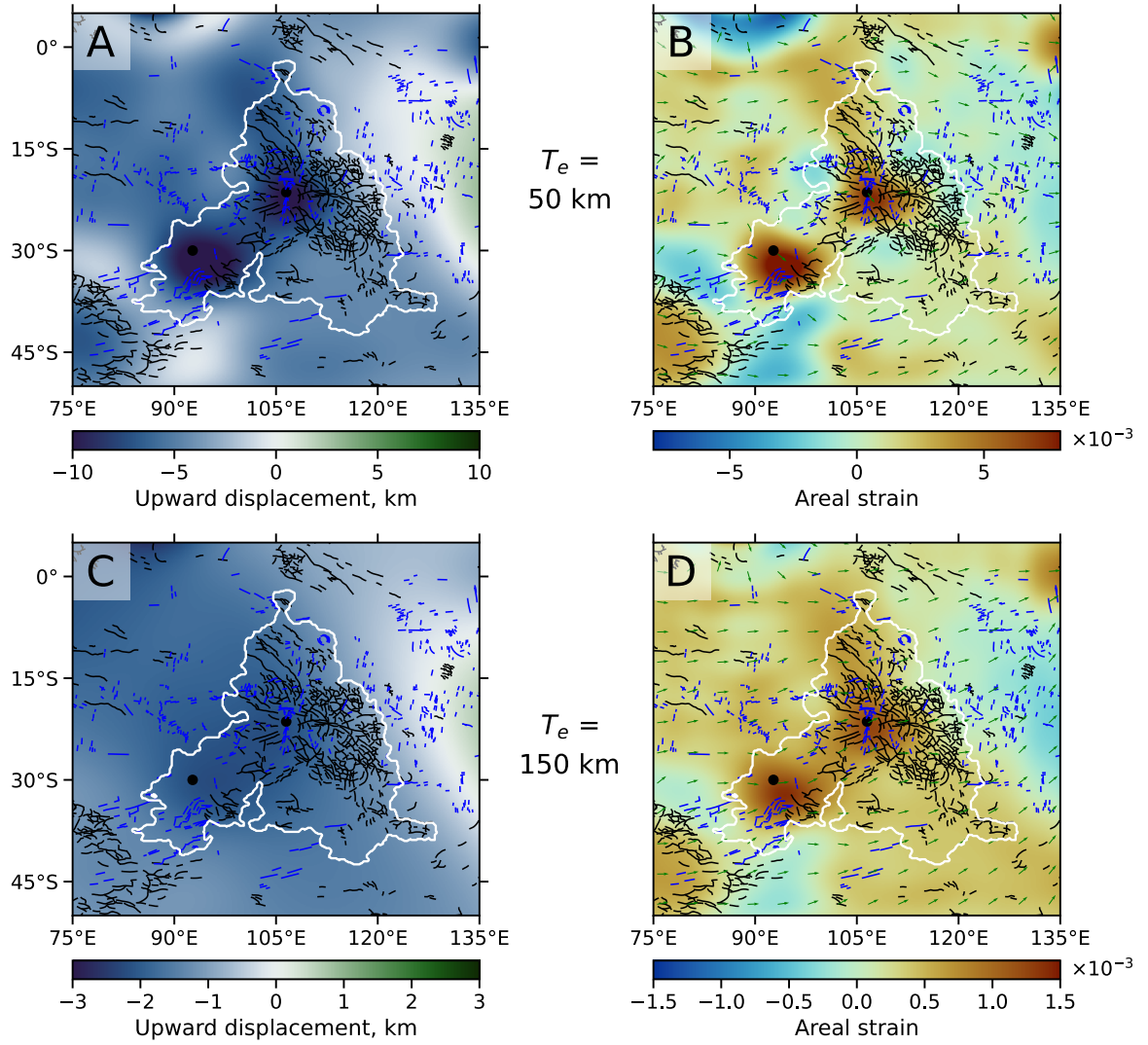


Figure 6 Displacement and strain at Hesperian Planum based on a global inversion of gravity and topography for two elastic thicknesses, 50 km (A, B) and 150 km (C, D). (left) Displacement of the lithosphere, where negative displacements correspond to downward flexure. (right) Predicted areal strain magnitude (color) and principal strain orientation (green arrows), where positive strains correspond to compression. Black dots indicate the location of Tyrrhena (106°E -21°N) and Hadriaca (93°E -30°N) Montes.

In addition to not fitting the observed strain magnitude, the inferred strain concentration is significantly different from that observed and this model cannot

explain the local strain concentration around 114°E 23°S, nor the inferred strain direction. The RMS difference between the observed normalized tectonic strain concentration and that predicted using elastic thicknesses of 50 and 150 km are about 0.29 and 0.31, respectively, and thus neither model provides a better match to the distribution of strain (these values are primarily useful for comparing models to one another; see below). The RMS difference between the observed azimuth of faulting and that modeled is 27.5° and 29.0° for elastic thicknesses of 50 and 150 km, respectively. For reference, if the wrinkle ridges were randomly arranged, the RMS difference would be about 26°, and thus the predicted azimuth of faulting in these models is incompatible with the observations. Together, these results suggest that the driving mechanism that generated significant compression on Hesperia Planum is not recorded in present-day gravity and topography.

5.3 Predictions of the mantle plume model

Finally, we consider the predictions of the dynamic model in which a sequence of plume uplift and subsidence is responsible for the observed tectonic strain. We first limit ourselves to the case of plume effects alone, not accounting for an additional contribution from the flood basalt loading. The fit to the plume head radius and location is based only on the tectonic strain concentration and fault orientations (Figure 7), and the other plume head and lithosphere parameters were obtained from fitting the strain magnitude within the peak strain area at Hesperia Planum after correcting for the subsequent global contraction (Figure S2).

To determine the plume head center position, we compared the predicted strain concentration and azimuth of faulting to the tectonic record for various input parameters and plume locations. The minimum misfit between the tectonic and modeled strain concentration is obtained when the plume is centered where wrinkle ridges cluster, near $114^{\circ}\text{E } 22^{\circ}\text{S}$ for plume head radii of 600 to 800 km (Figure 7, left). Placing the plume center within $\pm 2^{\circ}$ latitude and $\pm 4^{\circ}$ longitude of that location provides better fits to the strain concentration than did the global gravity and topography in the previous section, with RMS misfits of 0.17–0.22 versus 0.29. The misfit between the predicted azimuth of faulting and observed pattern of wrinkle ridges is also minimized for a plume centered near $114^{\circ}\text{E } 22^{\circ}\text{S}$ (Figure 7, right), and better fits the data for the plume model than our previous inversions, with a minimum value of 22.5° versus 27.5° . Additional low misfit solutions are found for plume heads centered close to the edge of the planum to the north-west, south-east, and south-west. These plume center locations are allowed because of the orthogonal pattern of wrinkle ridges, but are inconsistent with the strain concentration. Thus, the fits to those two independent quantities suggest a similar plume location around $114 \pm 4^{\circ}\text{E } 22 \pm 2^{\circ}\text{S}$, which is about 400 km east of Tyrrhena Mons. These tectonic analyses further show that Tyrrhena Mons is not the preferred plume head center, with higher misfits of ~ 0.30 and 27° , as a few ridges appear circumferential and radial about Tyrrhena Mons, but the majority show no clear relationship to the central volcano. Nevertheless, we acknowledge that some ridges could have formed following the evolutionary sequence and loading of

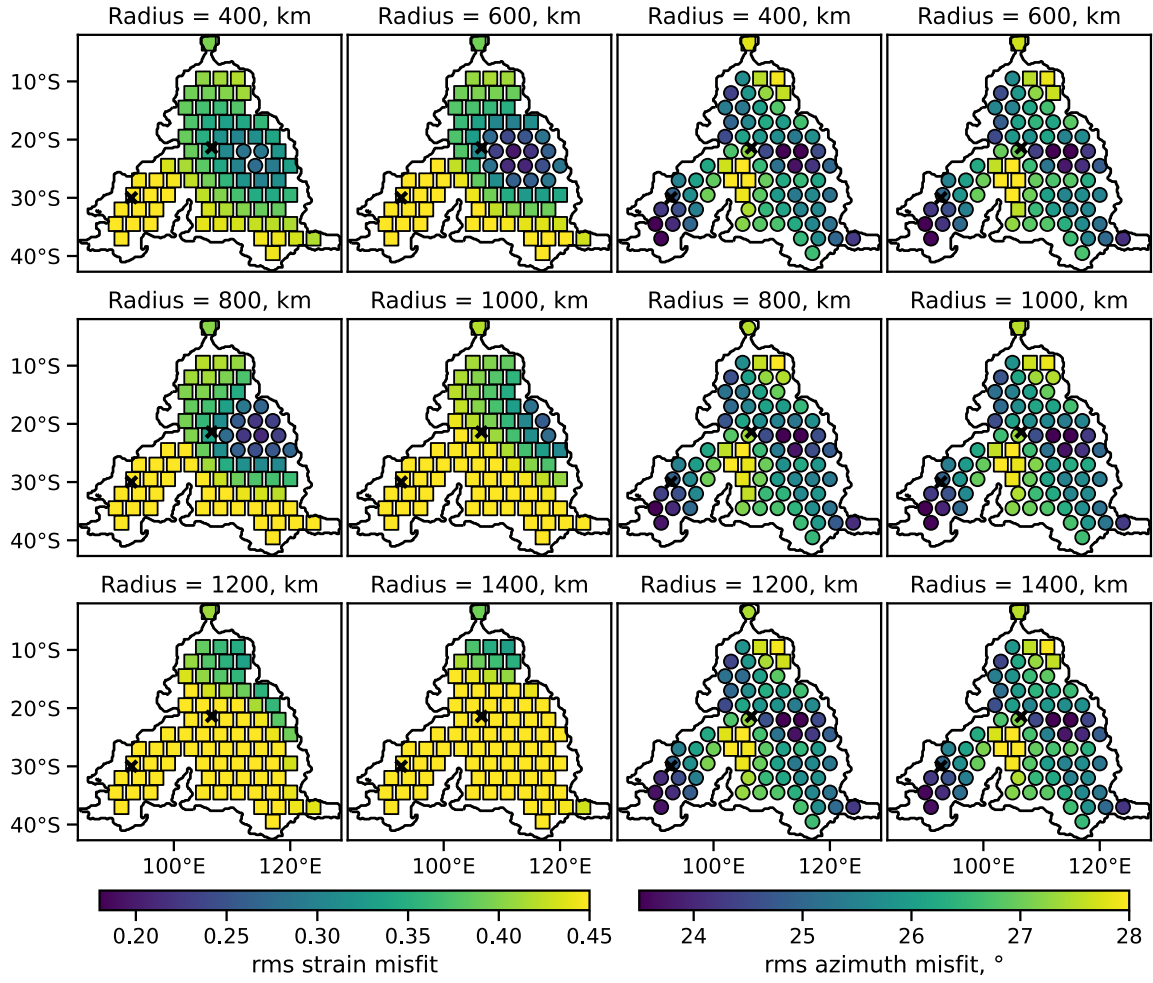


Figure 7 Plume head center at Hesperia Planum. Influence of the plume head radius on the RMS misfit between the observed and modeled strain concentration (2 leftmost columns) and azimuth (2 rightmost columns) as a function of the position the plume head (colored markers). The elastic thickness, plume head density and aspect ratio were set to 120 km, -35 kg m^{-3} , and 0.20 respectively. Where the RMS misfit is lower than from the global gravity and topography inversion (27.5°) location markers are dots, and are squares elsewhere. The black crosses indicate the position of Tyrrhena and Hadriaca Montes.

695 that shield volcano, as it is predicted to be associated with large compressive strains
 696 (Figure 6).

697 Using this same approach, we find similar best-fit plume locations for all
 698 parameters we investigated (Figure S3, S4, and S5), and we suggest that the paleo-
 699 plume center was located near $114 \pm 4^\circ\text{E}$ $22 \pm 2^\circ\text{S}$. Large elastic thicknesses ($>75 \text{ km}$)

700 better reproduce the observed strain concentration (Figure S3), but varying the density
701 contrast and aspect ratio of the plume head was found to not significantly influence the
702 predicted strain pattern (Figure S4 and S5). This is to be expected given that these two
703 parameters have a minor influence on the wavelength of deformation, which controls
704 the strain concentration and azimuth.

705 Next, we make use of the best-fit plume location and determine what parameters
706 best reproduce the observed strain magnitude (Figure S2, Text S3). For the models
707 with no flood lavas, the plume head density contrast should be at least -25 kg m^{-3}
708 (thermal anomaly of about 215 K) for an aspect ratio of less than 0.25, and up to -40
709 kg m^{-3} ($\sim 350 \text{ K}$) assuming an aspect ratio of 0.15. The minimum aspect ratio is
710 obtained for the largest density contrast investigated (-60 kg m^{-3}) and is 0.11, and any
711 elastic thickness can satisfy the data (Figure S2). For these parameters our predicted
712 uplifts range from 1 to 5 km. Importantly, though subject to large tradeoffs (Text S3),
713 these ranges of accepted parameters are consistent with plume head characteristics as
714 constrained on Earth (section 4.3), and imply that the observed tectonic record can be
715 explained solely by the dynamics of an Earth-like mantle plume. When accounting for
716 the presence of the flood basalt, with an initial thickness of 500 m, less strain is
717 required to originate from the plume head. The density contrast and aspect ratio are
718 both reduced, with minimum allowed values down to -10 kg m^{-3} and 0.09, respectively
719 (Figure S2).

720 In summary, the best-fit model has a plume head centered at $114^{\circ}\text{E } 22^{\circ}\text{S}$, with
721 a radius of 700 km and thickness of 280 km. The best-fit plume head density contrast

722 is -35 kg m^{-3} (Figure 8A) and is located underneath a 120-km thick elastic
723 lithosphere. This model predicts about 2.8 km of plume-induced uplift (B), which is
724 consistent with the terrestrial record (e.g., Sengör, 2001; Saunders et al., 2007). Post-
725 subsidence compressional strain is predicted to concentrate where wrinkle ridges
726 cluster (Figure 8C), and thus better explains the distribution of strain than the passive
727 volcanic loading scenario (Figure 6). Comparing the normalized modeled and
728 observed strain distributions gives an RMS misfit of 0.17, which is significantly
729 lower than found with passive volcanic loading only (0.29). In the restricted
730 investigated area, the mean strain value is about 1.61×10^{-3} , which compares
731 favorably to the observed value of $1.44\text{--}1.73 \times 10^{-3}$, where we have corrected the
732 latter for a global contractional strain of $1.47\text{--}1.76 \times 10^{-3}$ (section 3.2). The predicted
733 azimuth of faulting (Figure 8D) provides a better fit to the observed orthogonal and
734 short-wavelength strain pattern, with an RMS misfit of 22.5° , which is better than
735 for both a random distribution (26°) and the predictions from the passive model
736 (27.5°). That there is still a substantial mismatch in wrinkle ridge orientations

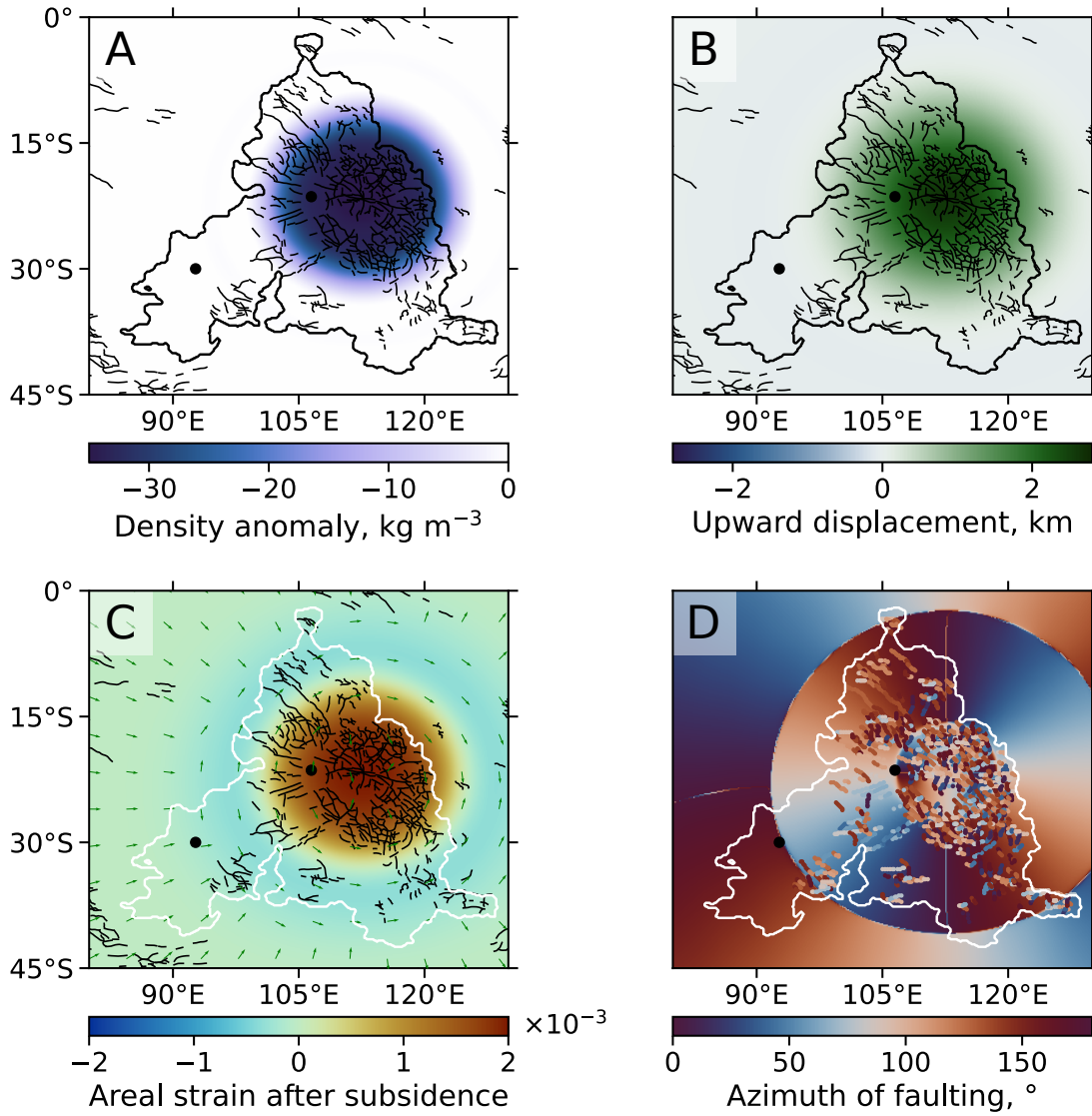


Figure 8 Best-fitting example of displacement and strain for a plume located at 114°E 22°S assuming a plume aspect ratio of 0.2 and an elastic thickness of 120 km. (A) The assumed density anomaly for a plume head radius of 700 km and density contrast of -35 kg m^{-3} , (B) the vertical displacement, (C) the areal strain magnitude after subsidence of the uplift (color) and maximum strain direction (green arrows), and (D) the predicted azimuth of faulting (background map) in comparison to observed wrinkle ridges (superposed lines).

737 indicates that other factors may also be influencing fault orientations, including pre-
 738 existing basement structures and other sources of non-isotropic strain. We note that

over much of the plume area, the difference between the two horizontal compressional strain magnitudes is small, and the orientation of faulting is not strongly controlled by these stresses.

6 Discussion – Implications of a mantle plume origin

6.1 Position of the mantle plume and relative formation sequence of Hesperia Planum

In its simplest sense, a plume model for the origin of Hesperia Planum would predict a transition from plume-induced uplift and flood basalt eruption to shield building, all focused on the plume center (Figure 2). In order to fit the observed tectonic record, we have constrained the plume head to have been located near 114°E 22°S when most of the deformation occurred. This position correlates with a cluster of wrinkle ridges and local crustal thinning (Figure 3B and C), but is about 400 km east of Tyrrhena Mons. Age relationships based on crater counting suggest that the ages of Tyrrhena Mons and Hesperia Planum overlap, with a major phase of shield volcanism postdating the lava plain formation (section 3.3). If the plume model is correct, it must be reconciled with the evolution of volcanism in both space and time.

Given the lack of plate tectonics on Mars, the offset between the plume center, the center of Hesperia Planum, the magnetic anomaly (see section 6.2), and Tyrrhena Mons may be explained by asymmetric plume flattening or plume drift driven by mantle flow patterns. Both have been observed in the terrestrial record (Camp, 1995; Jordan et al.,

2004, Tarduno et al., 2003; Konrad et al., 2018), and are predicted to occur in geodynamic models (e.g., Steinberger, 2000; Arnould et al., 2019). The simplest explanation for the temporal evolution would be that the bulk of Tyrrhena Mons is younger than the bulk of Hesperia Planum, in which case the volcanic shield formed from plume tail material after the main flood basalt sequence. Given uncertainties in the ages of both features, this scenario is consistent with observations.

6.2 Demagnetization and elastic thickness

Further insights into the elastic thickness and thermal anomaly associated with the plume head can be obtained by investigating the crustal magnetic field. The circular 400-km-diameter magnetic low, centered at 114°E 26°S, is located close to where wrinkle ridges cluster, within our expected plume head radius and only about 240 km away from the best-fit center. Based on its shape and position, it is possible that this magnetic low reflects the initial impingement of the hot mantle plume that partly thermally demagnetized the lithosphere before flattening and cooling underneath it. Laboratory experiments have constrained the flattened plume head to be generally about 3 times the initial spherical plume diameter (e.g., Griffiths & Campbell, 1991), and this would be consistent with our model favoring a final plume diameter of 1200 to 1600 km (3 to 4 times the size of the demagnetized region). Areas with decreased magnetism have also been shown to correlate with the presence of magmatic underplating, representing the paleo-plume location, in some terrestrial large igneous provinces (Blanchard et al., 2017; Ernst et al., 2019).

In order for the crust to be thermally demagnetized, most of the crust bearing the magnetic carrier should be raised above the Curie temperature, which is about 850–950 K considering hematite and magnetite as the most likely magnetic minerals of the crust (e.g., Dunlop & Arkani-Hamed, 2005). To first order, the 850–950 K isotherm is about the temperature at the base of the elastic lithosphere for a standard wet rheology (e.g., Plesa et al., 2018). This implies that the depth of magnetization is representative of the thickness of the elastic lithosphere. A localized power spectrum analysis found the depth of magnetization to be highest in the Hesperia Planum, at 50 to 90 km (Gong & Wieczorek, 2021). Given that the Curie temperature isotherm (roughly equal to the elastic thickness) must be deeper than the bounds on the depth of magnetization, the results of Gong and Wieczorek imply that the elastic thickness of the lithosphere is larger than 50 to 90 km, which is consistent with our preferred elastic thicknesses that are greater than 75 km (section 5.3). Similarly, the local demagnetization near the center of our preferred plume location indicates that the Curie isotherm and base of the lithosphere at this location were shallower at one point during the evolution of the planum, consistent with the thermal effects of a mantle plume.

7 Conclusions

The late Noachian to early Hesperian volcanic ridged plains represent a major episode in the geodynamic evolution of Mars, but little is known about the context of their formation. Flood basalts at Hesperia Planum have the highest density of

wrinkle ridges on the planet, suggesting a unique magmatic-tectonic evolution. Together with the planum's characteristic tectonic record, its local surface depression, bottom loading (crustal thinning or density increase), and magnetization reveal that Hesperia Planum was subject to a significant geologic event that heated and deformed the lithosphere. Our tectonic analysis has shown that the planum displays a clear local concentration of wrinkle ridges with a compressional strain of 3.20×10^{-3} on average and up to 4.75×10^{-3} . Both are significantly higher than on similarly aged surfaces ($\sim 1.47\text{--}1.76 \times 10^{-3}$) or than predicted by thermal evolution models ($\sim 2.0 \times 10^{-3}$, Hauck & Phillips, 2001; Andrews-Hanna et al. 2008). Using a flexural loading model based on present-day gravity and topography, we have demonstrated that the observed depression is not a result of an erosional event, and that the observed compression cannot be the result of passive volcanic loading.

We have explored the possibility that Hesperia Planum formed dynamically, as plume-induced flood basalts following an evolutionary path similar to that of some continental flood basalts on Earth (e.g., Richards et al., 1989; Campbell & Griffiths, 1990; Campbell, 2007). In this scenario, the surface would have recorded significant strains during a reversible sequence of uplift and subsidence, while the present-day gravity and topography would not preserve a record of this loading. The plume-induced uplift model is able to explain the observed peak strain magnitude, strain concentration, and wrinkle ridge orientations. The best-fit plume and lithosphere parameters are consistent with both terrestrial observations (White

824 & McKenzie, 1989; Saunders et al., 2007) and the seminal laboratory experiments
825 by Griffiths & Campbell (1990). Based on the fit to the tectonic record, the best-fit
826 paleo-plume head center was located around 114°E 22°S. The presence of a plume
827 at that location correlates with and provides a robust explanation for the presence
828 of a local circular magnetic low, where we favor a partial crustal demagnetization
829 associated with the initial spherical plume head before it flattened underneath the
830 lithosphere.

831 This study has used Hesperia Planum as an archetype for a larger class of
832 similar Noachian-Hesperian highland volcanic provinces, which includes Malea
833 Planum and Syrtis Major. These provinces formed during discrete geodynamic
834 episodes that were distinct from the contemporaneous volcanism that was largely
835 focused in the Tharsis region (Carr & Head, 2010). Our results suggest that the
836 Noachian-Hesperian martian mantle was not only characterized by a giant degree-
837 1 upwelling beneath Tharsis, but also by several smaller localized mantle plumes
838 with Earth-like characteristics. Our analysis of the formation sequence for Hesperia
839 Planum and its similarity to the evolution of large igneous provinces on Earth
840 supports the hypothesis that some highland volcanic provinces on Mars may
841 formed as a result of large plumes arising from the deep mantle.

842

Acknowledgments

The thin-shell flexural model can be found at Broquet (2022), and the main equations are given in Text S2. The spherical harmonic model of the gravity field is available at Zuber et al. (2010), and the color maps used in this study are from Crameri (2018). Some of this work made use of the SHTOOLS package of Wieczorek & Meschede (2018). A.B. kindly thanks the Laboratoire Lagrange of the Observatoire de la Côte d’Azur for hosting him during portions of this work, and A. Lasbordes for fruitful discussion. We thank Richard Ernst and an anonymous reviewer for their comments that helped improve the manuscript. This work was supported by grant 80NSSC17K0059 from the NASA Solar System Workings program to JCAH.

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