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1 **Different Martian Crustal Seismic Velocities across the Dichotomy**
2 **Boundary from Multi-Orbiting Surface Waves**

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25 **Abstract**

26 We have observed both minor-arc (R1) and major-arc (R2) Rayleigh waves for the largest marsquake (magnitude
27 of 4.7 ± 0.2) ever recorded. Along the R1 path (in the lowlands), inversion results show that a simple, two-layer
28 model with an interface located at 21 - 29 km and an upper crustal shear-wave velocity of 3.05 - 3.17 km/s can fit the
29 group velocity measurements. Along the R2 path, observations can be explained by upper crustal thickness models
30 constrained from gravity data and upper crustal shear-wave velocities of 2.61 - 3.27 km/s and 3.28 - 3.52 km/s in the
31 lowlands and highlands, respectively. The shear-wave velocity being faster in the highlands than in the lowlands
32 indicates the possible existence of sedimentary rocks, and relatively higher porosity in the lowlands.

33 **Plain Language Summary**

34 The largest marsquake ever recorded occurred recently and waves propagating at the surface, called surface waves,
35 have been observed. Owing to the relatively large magnitude (i.e., 4.7 ± 0.2) of this event, surface wave energy is
36 still clearly visible after one orbit around the red planet. The shortest path taken by the wave propagating between
37 the source and the receiver is located in the northern lowlands, near the boundary with the southern highlands (called
38 dichotomy). The surface wave traveling in the opposite direction, following the longer distance between the quake
39 and the seismic station, mostly passes through the highlands. Analyses of these two paths reveal that the average
40 shear-wave velocity is faster in the highlands than in the lowlands near the dichotomy boundary. This lower velocity
41 in the lowlands may be due to the presence of thick accumulations of sedimentary rocks and relatively higher porosity.

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43 **Key Points:**

- 44 ● Analyses of the minor-arc and major-arc Rayleigh waves reveal different Martian crustal structures across the
45 dichotomy boundary
- 46 ● The average shear-wave velocity is faster in the highlands than in the lowlands near the dichotomy boundary.
- 47 ● The lower shear-wave velocity in the lowlands may be due to the presence of sedimentary rocks and relatively
48 higher porosity.

1. Introduction

50 The InSight (Banerdt et al., 2020) mission landed a very broadband seismometer (Lognonné et al., 2019), on Mars
51 in Nov 2018 to better understand the structure, evolution, and differentiation of the Martian crust. In nearly three and
52 a half years, more than one thousand marsquakes and several impact events have been recorded (Clinton et al., 2021;
53 Garcia et al., 2022; Posiolova et al., 2022). So far, multiple body-wave signals have been clearly recorded (Clinton
54 et al., 2021), but observable surface-wave signals were only recently reported for two impact events (Kim et al.,
55 2022).

56 Due to the lack of surface wave observations, previous seismological studies of the Martian crust and the topmost
57 mantle were mainly based on body waves and ambient noise correlations (e.g., Lognonné et al., 2020; Knapmeyer-
58 Endrun et al., 2021; Compaire et al., 2021; Schimmel et al., 2021; Li et al, 2022a and 2022b). Since the ray paths of
59 the above-mentioned signals are nearly vertically incident beneath the seismometer, most of the sampled sub-surface
60 structures are limited to the vicinity of the InSight landing site (i.e., within a radius of ~ 50 km). Nevertheless, there
61 were attempts to constrain the crustal structures at other locations (e.g., Li et al., 2022c).

62 On the 1222nd Martian day (sol) of the InSight mission, the largest marsquake (S1222a) ever recorded occurred
63 with a magnitude of 4.7 ± 0.2 , which corresponds to energy five times greater than the second-largest event
64 (Kawamura et al. 2022; InSight Marsquake Service, 2022). For the first time, both Rayleigh and Love waves were
65 observed on a planet other than Earth (Kawamura et al. 2022; Beghein et al. 2022; Xu et al., 2022).

66 On a radially stratified, laterally homogeneous sphere, a surface wave travels along the great-circle path. The wave
67 that travels along the shorter path between the source and the receiver is named the minor-arc Rayleigh wave (R1).
68 The wave that travels in the other direction, along the longer path, is named the major-arc Rayleigh wave (R2). The
69 Rayleigh wave arrival is called R3 when the minor-arc wave circuits the globe and eventually reaches the station
70 again. On Earth, R2 and R3 can fill the gap of poorly covered areas in tomographic models (constrained by R1 alone)
71 and thus enhance the resolution (e.g., Trampert and Woodhouse, 1996). On Mars, the R2 and R3 observations are
72 very important since there is only one station and thus there are lots of gaps to cover.

73 To observe the R2 and R3 Rayleigh waves, which are of much smaller amplitude than R1, the minimum moment
74 magnitude was estimated to be $M_w 6.0$ for quiet stations on Earth (Panning et al., 2005). On Mars, the minimum
75 requirement for marsquakes is between $M_w 4.6$ and $M_w 5.3$, mainly due to Mars's smaller radius (Panning et al.,
76 2005). As predicted, the large magnitude of event S1222a enabled observations of R2 and R3 on Mars (Panning et
77 al., 2022).

78 The epicentral distance of event S1222a is 37.0 ± 1.6 degrees, and the back azimuth is estimated to be between 96
79 and 112 degrees (Kawamura et al. 2022), placing the location of S1222a in the lowlands and near the dichotomy
80 boundary. The dichotomy boundary separates the northern lowlands and southern highlands and has highly debated
81 origins (Zhong and Zuber, 2001; Marinova et al., 2008; Andrews-Hanna et al., 2008; Frey, 2006). The R2 and R3
82 Rayleigh waves pass through both the lowlands and the highlands (Fig. 1), which are far from the InSight lander and
83 were scarcely sampled by seismic waves with the previously recorded marsquakes. These observations allow us to
84 put new constraints on the crustal structure away from the lander in both the lowlands and the highlands.

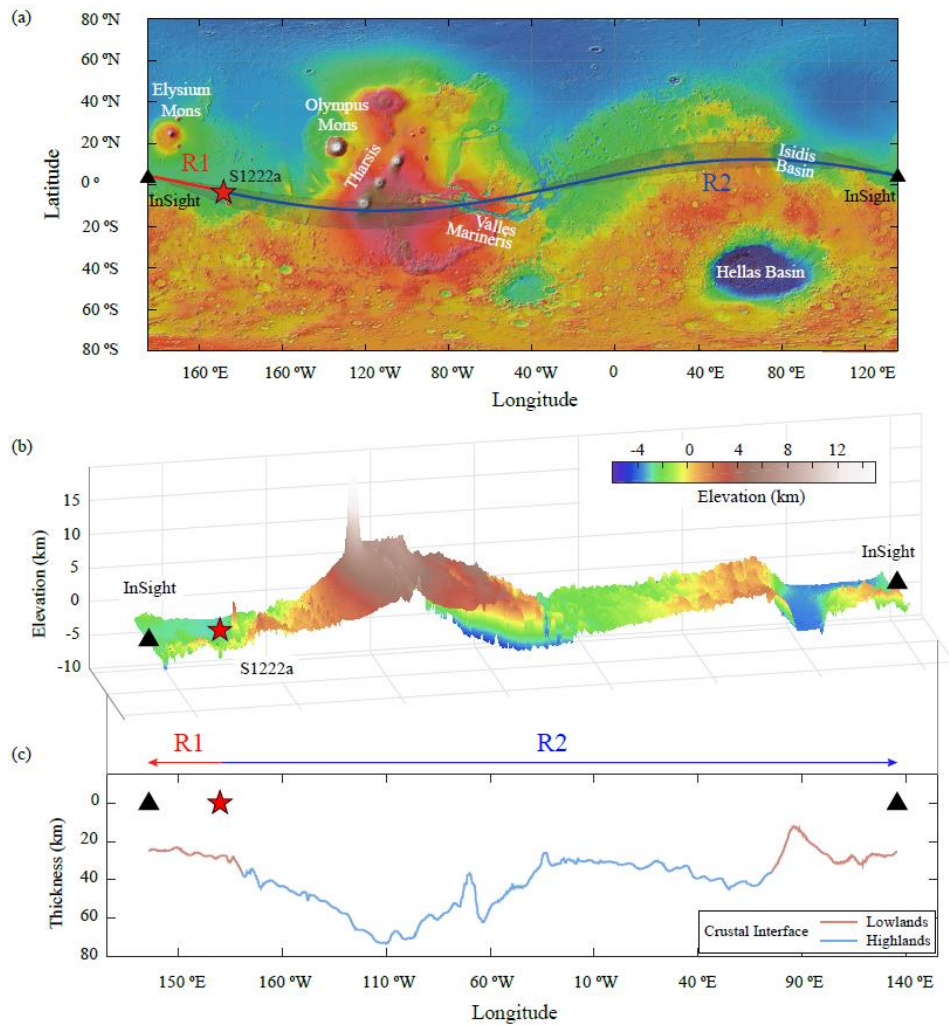
85 In this paper, we first measured the group velocity of the Rayleigh waves along the R1, R2, and R3 paths. Then,
86 we performed inversions for R1 and R2 (the R3 measurements are less reliable, as explained below) to derive seismic
87 velocity models that could explain the observations. After comparing the seismic velocity profiles in the lowlands
88 and highlands, we discuss the possible effects of the crustal variations on the observed group velocities of the minor-
89 and major-arc Rayleigh waves.

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96 **Figure 1.**

97 (a) MOLA (Mars Orbiter Laser Altimeter, Smith et al., 2001) map. The red star marks the optimal hypocenter of
 98 event S1222a (Kawamura et al., 2022), and the black triangle indicates the broadband seismometer of the InSight
 99 mission. The red and blue curves denote the great-circle paths of R1 and R2, respectively. The shaded grey areas
 100 illustrate all possible paths for R1 and R2 based on the back azimuth uncertainty of the event location (Kawamura
 101 et al., 2022).

102 (b) Elevation variations along the Rayleigh wave paths shown in (a). Regions within 10 degrees of the optimal great-
 103 circle path are displayed to account for the finite frequency effect (e.g., Zhou et al., 2004).

104 (c) Lateral variations of the upper crust along the Rayleigh wave great-circle path in (a). The depth of the interface
 105 at the InSight lander site is fixed at 26 km from the R1 inversion (in Fig. 3c), and the relative variations are
 106 constrained by gravity data (Wieczorek et al., 2022). The location of the dichotomy boundary is from Tanaka et
 107 al. (2014).

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112 **2. Data**

113 The minor arc Rayleigh wave, which travels $2,193 \pm 95$ km (using an equatorial radius of 3396.2 km), arrives
114 around 800 s after the origin time on the vertical component of the seismogram and can be observed over a relatively
115 broad frequency band on both the vertical (i.e., 10 to 60 s, in Fig. 2a) and the radial components (Fig. S1). Since the
116 seismometer is in an extremely harsh environment, the diurnal temperature variations could generate one-sided pulses
117 (hereafter referred to as glitches) that can contaminate the seismic signals. A comparison between the raw data and
118 the deglitched data (Scholz et al., 2020) shows that there is no significant glitch near the R1 arrival time (Fig. S1).

119 R2 travels a much longer distance of about $19,146 \mp 95$ km and arrives around 6,650 s after the origin time (Fig.
120 2b). R2 is observed within a relatively narrow frequency range (i.e., 25 to 35 s), and the amplitude is only about one-
121 fourth of that of R1. Although there is no glitch near the R2 arrival, another signal of a slightly smaller amplitude
122 occurs at about 6,800 to 7,000 s. Panning et al. (2022) excluded this signal as being the R2 phase since this later
123 arrival does not show strong and consistent elliptical particle motions at different frequency bands (which is expected
124 for the Rayleigh wave), and the corresponding marsquake location would be highly inconsistent with the one
125 constrained by body waves (Kawamura et al. 2022). Therefore, we prefer selecting the R2 signal at around 6,650 s.

126 The R3 wave, which travels $23,532 \pm 95$ km, overlaps with a glitch in the raw data (Fig. 2b). R3 is still visible on
127 the deglitched data and exhibits arrival times of about 8,100 s. However, since the deglitching procedure may not
128 perfectly fit the real glitch and thus could also remove some seismic signals, the R3 signals (e.g., the arrival time and
129 amplitude) are less reliable than R1 or R2. Therefore, in the following analysis, we mainly focus on R1 and R2 and
130 consider R3 as auxiliary data for benchmarking purposes.

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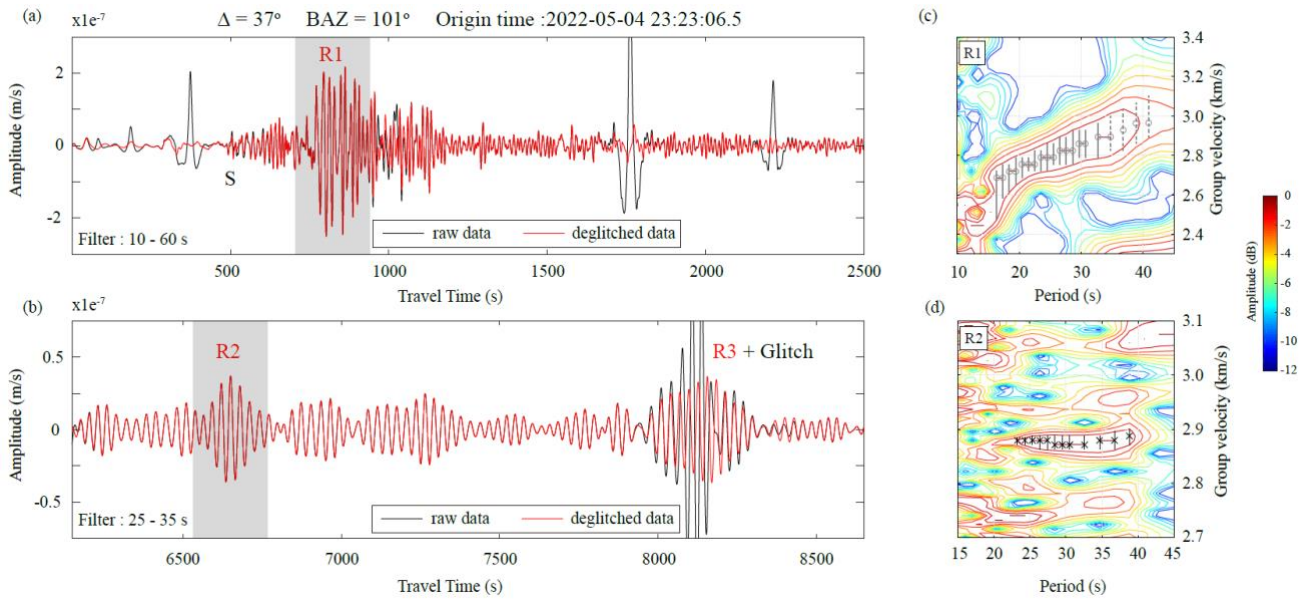
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142 **Figure 2.**

143 (a) Seismic waveforms (after removing the instrument response and bandpass filter from 10 to 60 s) on the vertical
 144 component near the arrival times of the S wave and R1. The black and red waveforms are the velocity records
 145 from the raw dataset and deglitched dataset, respectively. The shaded gray region marks the time window in
 146 which we performed the multiple filter analysis in (c).

147 (b) Waveforms on the vertical component near the arrivals of R2 and R3 with a filter band of 25 to 35 s.

148 (c) Multiple filter analysis (Herrmann, 2013) for R1 on the vertical component. The x- and y-axes are the period and
 149 group velocity, respectively. Color scales represent the energy contours (in dB) of the waveform envelope. The
 150 black circles indicate the preferred group velocity measurements. The vertical grey error bar spans 1 dB. Note
 151 that at periods greater than 32 s, the R1 measurements show discrepancies between the vertical and radial
 152 components (see Fig. S2) and thus are less reliable (denoted by the dashed error bars).

153 (d) Similar to (c) but for R2 on the vertical component.

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161 **3. Method and Results**

162 **3.1 Group Velocity Measurements**

163 To measure the Rayleigh wave group velocity, we first performed a multiple filter analysis (Herrmann, 2013) to
164 calculate the energy of the waveform envelope at different periods (Fig. 2c and d). Then, we selected the optimal
165 group velocity at each period corresponding to the maximum energy (in dB) of the waveform envelope. Finally, we
166 defined the uncertainty of the measurements using a 1dB threshold (Beghein et al., 2010), i.e., group velocities with
167 amplitudes within 1dB of the maximum amplitude are viewed as acceptable measurements.

168 The reliable period range for the R1 signal on the vertical component is from 17 s to 32 s, where the group velocities
169 are consistent with those measured on the radial component (Fig. S2). For the average group velocity along the R1
170 path, there are clear signs of positive dispersion, i.e., the group velocity increases with the period. Given the origin
171 time of 2022-05-04 23:23:06.5 (UTC) and the epicentral distance of 37 degrees (Kawamura et al. 2022), the R1 group
172 velocity increases from 2.70 ± 0.05 km/s at 17 s to 2.95 ± 0.05 km/s at 41 s (Fig. 2c, see Text S1 for discussions on
173 the location error).

174 We applied the same analysis to the R2 signal and found that it exhibits little dispersion between periods of 23 s
175 and 39 s, with a nearly constant group velocity of 2.90 km/s \pm 0.01 km/s (Fig. 2d). It is worth mentioning that the
176 uncertainty of the R2 group velocity is much smaller (less than one-fifth) than that of R1 due to its longer travel time
177 (See Text S1 for details).

178 Fig. 2b shows that R3 is significantly affected by a glitch. Nevertheless, based on the deglitched data, the multiple
179 filter analysis shows little dispersion and a nearly constant group velocity of about 2.90 km/s \pm 0.02 km/s (Fig. S3).
180 This similar value of the R2 and R3 group velocities confirms the reliability of the R2 measurements given their
181 similar paths (i.e., they share the same path for about 19,000 km or about 80% of the R3 path length).

182 **3.2 Inversion Along the R1 Path**

183 We inverted the group velocity measurements to constrain the average seismic velocity profile along the R1 path
184 using a Niche Genetic Algorithm (Koper et al., 1999; Li et al., 2021) to sample different possible models and the CPS
185 code package (Herrmann, 2013) to calculate the group velocities predicted by these models at different periods.

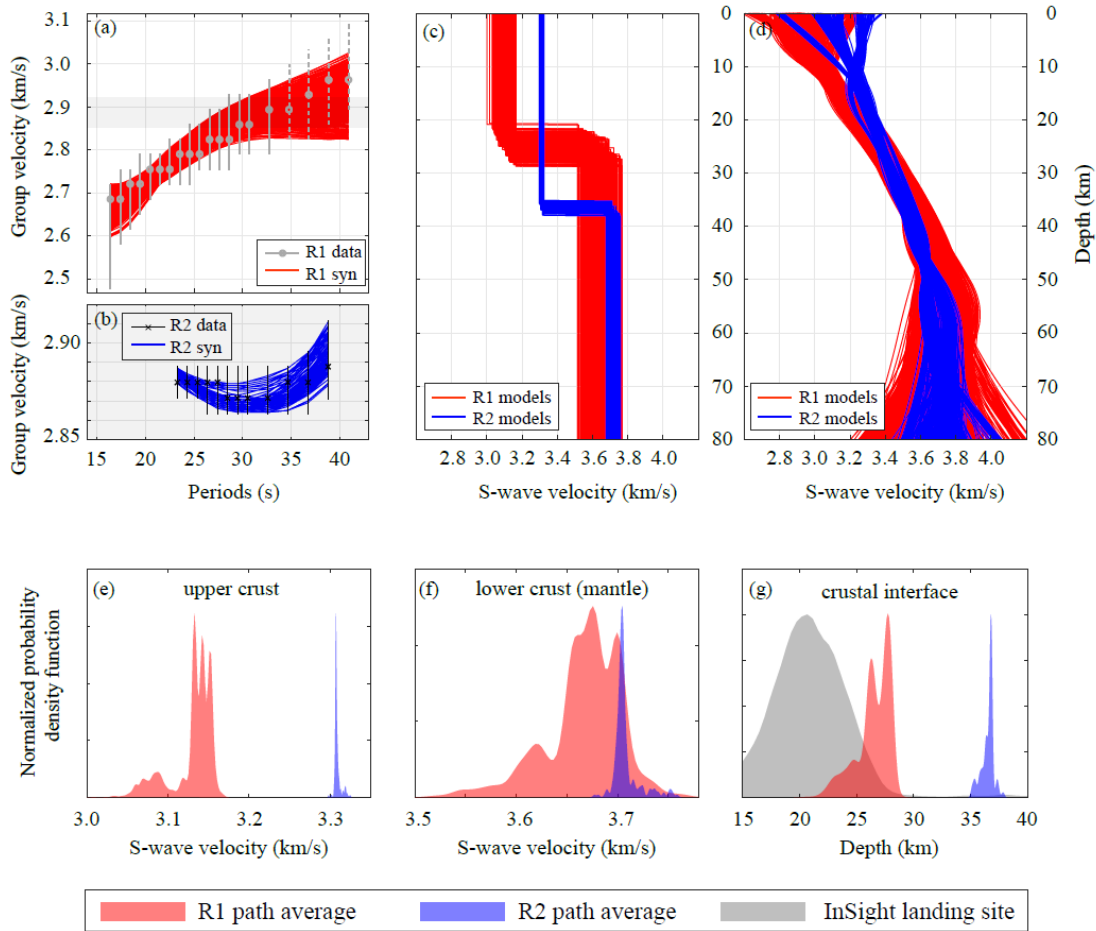
186 We chose to represent the model with a layered crustal parameterization because previous receiver function studies

187 (Lognonné et al., 2020; Kim et al., 2021; Knapmeyer-Endrun et al., 2021) exhibited the presence of sharp interfaces
188 in the crust. In addition, we preferred to set the number of layers to two because this model setting with only three
189 free parameters (i.e., the depth of the interface, and the shear-wave velocities above and below this interface) is less
190 likely prone to overfitting the data. If this simplified model fails to fit the observed group velocities, we can then
191 consider a more complex model. Based on the sensitivity curves of the Rayleigh waves at periods shorter than 40 s
192 (Fig. S4), we only constrained the structure above 80 km depth. At greater depths, we used the values of the
193 InSight_KKS21GP model (Knapmeyer-Endrun et al., 2021; Khan et al., 2021; Stähler et al., 2021a and 2021b), which
194 was constructed from teleseismic body waves (See Text S2 for the model setup). We also tested a smoothed model
195 setting with eight third-order B-splines (Fig. 3d) and found that the average velocities above (or below) the interface
196 are similar between the discontinuity and the smoothed model settings (Fig. S5).

197 Fig. 3c shows that a simplified model with two layers in the top 80 km can reproduce the R1 dispersion curve.
198 Specifically, the shear-wave velocity is estimated to be 3.05 - 3.17 km/s in the top layer, and 3.50 - 3.78 km/s in the
199 bottom layer. The interface depth estimation is 21 - 29 km. Fig. 3f shows that the depth of the inverted interface along
200 the R1 path overlaps with the depth of the intra-crustal interface beneath the InSight lander constrained by the receiver
201 function (Knapmeyer-Endrun et al., 2021), although the mean value deviates (possibly due to the difference in the
202 regions to which those two methods are sensitive).

203 We note that Knapmeyer-Endrun et al. (2021) provided two possible crustal thicknesses beneath the lander: a
204 thinner crustal model with a shallow Moho depth of 20 ± 5 km, and a thicker crustal model (where the interface at
205 around 20 km is interpreted as an intra-crustal interface) with a deeper Moho depth of 39 ± 8 km (though with a
206 smaller impedance contrast across this deeper Moho). Here, we prefer to refer to the interface at 21- 29 km (inverted
207 in this study) as the intra-crustal interface since subsequent studies from Kim et al. (2021) and Durán et al. (2022)
208 favored the thicker crustal scenario. We also want to note that we cannot resolve the deeper Moho (if it exists) since
209 only the discontinuity with the most significant velocity jump will be detected with the current one-interface setting.

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213 **Figure 3.**

- 214 (a) Group velocity measurements (grey circles), uncertainties (vertical bars), and synthetic group velocity dispersion
 215 curves (in red, corresponding to the models shown in (c) with the same color) along the R1 path. The range of
 216 the y-axis (i.e., group velocity) in the shaded gray region is the same as that in (b).
 217 (b) Similar to (a), but for the R2 path, and the corresponding models (in blue) shown in (c).
 218 (c) Shear-wave velocity profiles for all the acceptable models along the R1 (in red) and R2 (in blue) paths from the
 219 inversion for the discontinuity model setting.
 220 (d) Similar to (c), but for the smoothed (i.e., third-order B-splines) model setting.
 221 (e) Normalized probability density functions of the shear-wave velocities in the upper crust for all the acceptable
 222 models (R1 models in red, and R2 models in blue) in (c).
 223 (f) Similar to (e), but for the shear-wave velocities in the lower crust and mantle (i.e., down to 80 km).
 224 (g) Similar to (e), but for the depth of the crustal interface. The probability density function in gray shows the depth
 225 of the base of the intra-crustal discontinuity beneath the lander from the receiver function study (Knappmeyer-
 226 Endrun et al., 2021).

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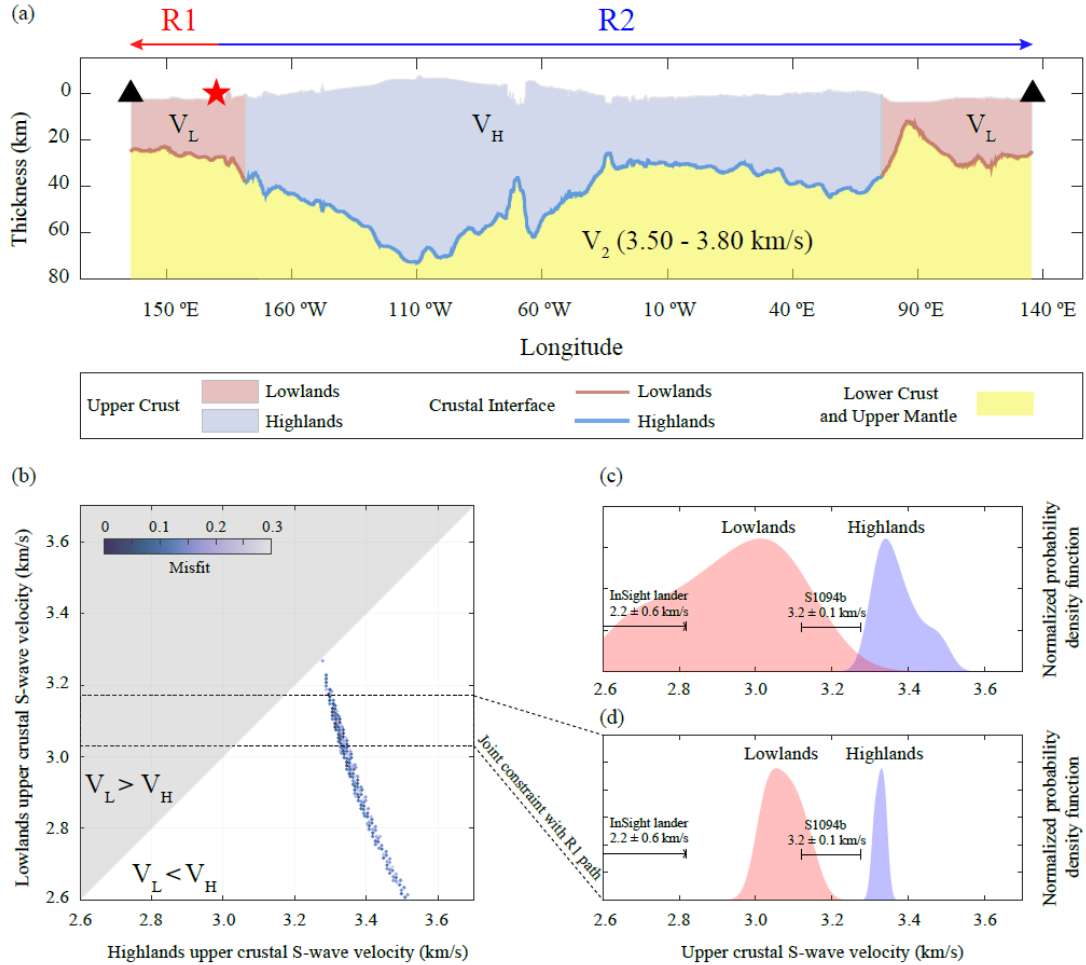
230 **3.3 Inversion Along the R2 Path**

231 We note that it is reasonable to perform the inversion with a representative one-layer model along the entirety of
232 the R1 path since there are few topographic changes (e.g., Fig. 4a). However, along the R2 path, it may be preferable
233 to account for the significant lateral variations in both topography and crustal thickness (Wieczorek et al., 2022),
234 rather than simply using an average one-layer model (e.g., Fig. 3c) to explain the R2 dispersion curve in Fig. 3b.

235 We thus divided the R2 path into 3,230 sub-paths, and within each sub-path (of which the length is shorter than
236 the wavelength) we can assume a one-layer model with a similar setup as for R1. The upper crustal thickness models
237 along the R2 path are modified from Wieczorek et al. (2022) as prior information (See Text S3 for the R2 model
238 setup). There are only three free parameters in the R2 inversion, i.e., the upper crustal shear-wave velocities in the
239 lowlands (V_L) and highlands (V_H), and the shear-wave velocity below the crustal interface (V_2).

240 Grid search results show that within the given range of V_2 (i.e., 3.50 - 3.78 km/s from the R1 inversion results in
241 Fig. 3e), a series of models (Fig. 4b), with upper crustal velocities of 2.61 - 3.27 km/s in the lowlands and 3.28 - 3.52
242 km/s in the highlands, can fit the R2 measurements (Fig. S6).

243 The probability density functions in Fig. 4c illustrate that the upper crustal shear-wave velocity in the highlands
244 (i.e., 3.28 - 3.52 km/s) is systematically faster than that in the lowlands (i.e., 2.61 - 3.27 km/s). We further reduced
245 the number of acceptable models from the R2 inversion with the extra constraint from the R1 inversion (i.e., upper
246 crustal velocity in the lowlands should be in the range of 3.05 - 3.17 km/s). The down-selected models, which satisfy
247 both the R1 and R2 observations, also demonstrate a greater upper crustal shear-wave velocity in the highlands (i.e.,
248 3.17 - 3.33 km/s) than that in the lowlands (i.e., 3.05 - 3.17 km/s).



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251 **Figure 4.**

252 (a) Model setup for the R2 inversion. The upper crustal thickness models modified from Wieczorek et al. (2022) are
 253 used as prior information and fixed in the inversion. Along the R2 path, the only three free parameters are the
 254 upper crustal shear-wave velocities in the lowlands (V_L) and highlands (V_H), and the shear-wave velocity (V_2)
 255 below the crustal interface (See Text S3 for the R2 model setup).

256 (b) Grid search results for the upper crustal shear-wave velocity in the lowlands (V_L , the y-axis), and the upper
 257 crustal shear-wave velocity in the highlands (V_H , the x-axis) along the R2 path. The acceptable models that can
 258 fit the R2 measurements are shown as filled circles, with color scales corresponding to the misfits (See Text S2
 259 for the misfit function). The dashed region indicates where the models also satisfy the R1 observations (i.e., an
 260 upper crustal shear-wave velocity of 3.05 - 3.17 km/s in the lowlands). The shaded gray area to the left of the
 261 diagonal indicates the region where the upper crustal velocity in the lowlands is faster than that in the highlands.
 262 Note that at each grid point, a grid search for the shear-wave velocity below the crustal interface (V_2) is also
 263 performed.

264 (c) Normalized probability functions of the upper crustal shear-wave velocities (lowlands models are in red, and
 265 highlands models are in blue) for the acceptable models in (b). S-wave velocities beneath the InSight lander (in
 266 the depth range of ~8 - 20 km, Knappmeyer-Endrun et al., 2021) and along the great circle path from the meteorite
 267 impact S1094b (Kim et al., 2022) are also annotated.

268 (d) Similar to (c), but for the models that satisfy both the R1 and R2 observations (i.e., in the dashed region in (b)).

269

270 **4. Discussion**

271 **4.1 Comparison with the InSight landing site**

272 The most sensitive depth range of the Rayleigh waves analyzed is from ~ 10 to 50 km (Fig. S7), though there are
273 also a few sensitivities to greater depths. Therefore, we have better constraints on the upper crust (i.e., above the
274 crustal interface), than the lower crust and mantle.

275 Since Lognonné et al. (2020) and Knapmeyer-Endrun et al. (2021) obtained the first Martian crustal structure
276 beneath the InSight landing site, whether it represents local geologies or global features has been debated. In the
277 similar depth range of $\sim 8 - 20$ km, beneath the lander, the S-wave velocity is estimated to be 2.2 ± 0.6 km/s
278 (Knapmeyer-Endrun et al., 2021). This upper crustal velocity beneath the lander (also located in the lowlands) is
279 significantly smaller than the average velocity (i.e., ~ 3 km/s, see Fig. 4c and d) along the surface wave path in the
280 lowlands.

281 Similar S-wave velocity (i.e., ~ 3.2 km/s) of the upper crust in the lowlands (along the surface wave path from a
282 meteorite impact) has also been recently reported (Kim et al., 2022). This faster velocity has been attributed to the
283 different compositions and/or reduced porosity in the volcanic areas along the path (Kim et al., 2022).

284 For marsquake S1222a, the surface wave path also likely traverses the volcanic units in the lowlands (Fig. S8).
285 Therefore, the crustal porosity may have been occluded or obscured by later magmatism (e.g., Wilson and Head,
286 1994), resulting in a faster S-wave velocity than beneath the lander (where there are higher porosities).

287 **4.2 Upper Crust in the Lowlands and the Highlands**

288 The upper crust velocity in the lowlands, though larger than the InSight landing site, is systematically smaller than
289 that in the highlands (Fig. 4c and d). Possible interpretations for the smaller upper crustal shear-wave velocity values
290 in the lowlands than in the highlands could include some combination of differing compositions, reductions in
291 porosity at depth, and/or the presence of sedimentary rocks.

292 Variations in lithology and rock compositions are a common major cause of regional variations in crustal seismic
293 structure (Rudnick and Fountain, 1995). The Martian crust is thought to be of basaltic composition on average (Taylor
294 and McLennan, 2009). Recently, using global crustal thickness and thermal evolution models, Thiriet et al. (2018)
295 suggested that felsic rocks were more common in the Noachian-dominated southern highlands, leading to an average

296 bulk density of 480 kg/m^3 less than the northern lowland crust. However, this density difference should lead to lower,
297 rather than higher, seismic velocities (for a wide range of lithologies, e.g., Brocher, 2005) in the highlands.

298 Compositional differences could also affect the thermal regime and thus have an indirect influence on seismic
299 velocities. For example, a more differentiated southern highland crust would also be characterized by higher
300 abundances of heat-producing elements (Thiriet et al., 2018; Wieczorek et al., 2022) and, when coupled with a thicker
301 crust, higher heat flow (e.g., Hahn et al., 2011; Parro et al., 2017). A distinctive crustal thermal regime could lead to
302 very different depth-dependent porosity behavior.

303 Porosity can also significantly affect the seismic wave speed such that higher porosity results in lower shear-wave
304 velocity (e.g., Manga and Wright 2021). Both compaction and viscous deformation will result in porosity reduction
305 with depth. Gyalay et al. (2020) showed that the closure of pore space should occur over a narrow depth range of a
306 few kilometers and all porosity should have been removed for depths greater than about 12-23 km (Wieczorek et al.
307 2022). Accordingly, in the highlands, where the upper crust is thicker on average (Fig. 4a) and surface heat flow is
308 greater (Hahn et al., 2011; Parro et al., 2017), a larger proportion of the upper crust may lie beneath depths where
309 porosity has been lost, leading to shear-wave velocities that are higher than in the northern lowlands when averaged
310 over the thickness of the upper crust interrogated by the R2 paths.

311 To quantitatively assess the influence of porosity, we assumed that the velocity difference between the lowlands
312 and highlands is affected by porosity alone. With a pore closure depth of 23 km (Wieczorek et al. 2022) and an aspect
313 ratio of 0.1 (Heap, 2019), the velocity difference can be explained by a porosity reduction of 8 - 15% (Fig. S9, see
314 Text S4 for details).

315 The presence of thick accumulations of sedimentary rocks could also lead to lower density (and thus, for a wide
316 range of lithologies, low shear-wave velocities, e.g., Brocher, 2005) due to a variety of features/processes, including
317 the presence of elevated primary porosity for poorly consolidated and/or weakly cemented sedimentary rocks;
318 secondary porosity for diagenetically altered sedimentary rocks; low-density hydrated alteration phases, such as
319 phyllosilicates and amorphous silica, formed during chemical weathering and/or diagenesis (McLennan & Grotzinger,
320 2008; McLennan et al., 2019). However, for most locations on Mars, sedimentary rock thicknesses are on the order
321 of a few kilometers at most (e.g., Grotzinger et al., 2015; Grotzinger and Milliken, 2012; Bradley et al., 2002).
322 Therefore, based on our current knowledge of the total sedimentary mass and its distribution on Mars (e.g., McLennan,
323 2012, Grotzinger and Milliken, 2012), the presence of sedimentary rocks alone probably cannot explain the lower
324 upper crustal shear-wave velocity in the lowlands, where the crustal interface is located at $\sim 26 \text{ km}$.

325 **4.3 R2 Group Velocity**

326 Another major difference between the group velocity along R1 and R2 is that there is positive dispersion along R1
327 but there is almost no dispersion along R2. We found that for a typical velocity model derived for the highlands,
328 where the upper crustal thickness is greater than in the lowlands, the group velocity exhibits negative dispersion.
329 Therefore, the negative dispersion in the highlands (Fig. S7e) could cancel out the positive dispersion in the lowlands
330 (Fig. S7d), yielding little dispersion along the R2 path.

331 However, it is worth noting that long-period pressure-induced signals are also present in seismic data (Kenda et
332 al., 2020; Stutzmann et al., 2021) and are carried by the environmental wind, resulting in an apparent burst of noise
333 without dispersion. Given the relatively low signal-to-noise ratio for R2 and R3, this absence of dispersion could also
334 be an artifact of the wind noise.

335 **5. Conclusion**

336 We have observed two minor-arcs (R1, and R3) and one major-arc (R2) Rayleigh wave generated by the largest
337 marsquake ever recorded. The group velocities along the R1 path increase from 2.70 ± 0.05 km/s to 2.95 ± 0.05 km/s
338 between periods of 17 s and 42 s and can be explained by a two-layer model with a crustal interface depth of 21 - 29
339 km and an upper crustal shear-wave velocity of 3.05 - 3.17 km/s in the lowlands. The group velocity along the R2
340 path exhibits a nearly constant value of 2.90 km/s and can be explained by upper crustal thickness models constrained
341 from gravity data and upper crustal shear-wave velocities of 2.61 - 3.27 km/s and 3.28 - 3.52 km/s in the lowlands
342 and highlands, respectively. The lower shear-wave velocity in the lowlands compared to that in the highlands may be
343 due to the presence of thick accumulations of sedimentary rocks, and relatively higher porosity.

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352 This is InSight Contribution Number 278. InSight seismic data presented here
353 (http://dx.doi.org/10.18715/SEIS.INSIGHT.XB_2016) are publicly available through the Planetary Data System
354 (PDS) Geosciences node (InSight SEIS Data Bundle, 2021), the Incorporated Research Institutions for Seismology
355 (IRIS) Data Management Center under network code XB and through the Data center of Institut de Physique du
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358 providing SEED SEIS data.

359 **Conflict of Interest**

360 The authors declare no conflicts of interest relevant to this study.

361 **Open Research**

362 The datasets used in this study are achieved and released by InSight Mars SEIS Data Service and are
363 available to the science community (InSight Mars SEIS Data Service, 2019; InSight Marsquake Service,
364 2022). Data sets (both the raw data and the deglitched data in SAC format, after removing the instrument
365 response) for this research are available on the Zenodo repository: <https://doi.org/10.5281/zenodo.7325555>.
366 The InSight_KKS21GP model is available on the IPGP Data Center (doi: 10.18715/IPGP.2021.kpmqrmz8).

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