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Structure Along the Martian Dichotomy Constrained by Rayleigh and Love Waves and their Overtones

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28 Abstract

29 Using seismic recordings of event S1222a, we measure dispersion curves of Rayleigh and Love
30 waves, including their first overtones, and invert these for shear velocity (V_s) and radial anisotropic
31 structure of the martian crust. The crustal structure along the topographic dichotomy is
32 characterized by a fairly uniform vertically-polarized shear velocity (V_{sv}) of 3.17 km/s between
33 ~5-30 km depth, compatible with the previous study by Kim et al. (2022). Radial anisotropy as
34 large as 12 % ($V_{SH} > V_{SV}$) is required in the crust between 5-40 km depth. At greater depths, we
35 observe a large discontinuity near 63 ± 10 km, below which V_{sv} reaches 4.1 km/s. We interpret
36 this velocity increase as the crust-mantle boundary along the path. Combined gravimetric modeling
37 suggests that the observed average crustal thickness favors the absence of large-scale density
38 differences across the topographic dichotomy.

39 Plain Language Summary

40 The first detection and analysis of surface waves on Mars (Kim et al., 2022) revealed that the
41 crustal structure away from the *InSight* lander is fairly uniform between 5-30 km depth in the
42 northern lowlands. This is strikingly different from the crustal structure inferred beneath the lander.
43 The largest marsquake recorded during the *InSight* mission to Mars, S1222a, provides the first
44 clear signals of both types of surface waves – called Rayleigh and Love waves – as well as their
45 first overtones. We analyze the speed at which these waves travel changes with their frequency to
46 see deeper into Mars than possible with previous data. We find that the crustal structure along the
47 path to S1222a, which covers a different part of the northern lowlands, is similar to that found
48 previously, suggesting that uniform velocities in the 5-30 km depth may be characteristic for this
49 region. By combining our seismic data with variations in the strength of gravity, we determine that
50 the density of the crust in the northern lowlands and the southern highlands is similar. Finally, by
51 analyzing both types of surface waves, we find that the speed of horizontally-polarized waves is
52 up to 12% faster than that of vertically-polarized waves.

53 1 Introduction

54 The interior of Mars has been probed by seismic waves since the *InSight* mission (Banerdt et al.,
55 2020) placed a very sensitive, three-component broadband seismometer (Lognonné et al., 2019)
56 on its surface to monitor the planet's ground vibrations. Since then, analysis of waveforms of body
57 wave phases from marsquakes (Giardini et al., 2020) and impacts (Garcia et al., 2022; Posiolova
58 et al., 2022) have resulted in important discoveries about its interior structure. They include the
59 characterization of crustal structure (Lognonne et al., 2020; Knapmeyer-Endrun et al., 2021; Kim
60 et al., 2021a) and mantle velocity structure (Khan et al., 2021; Drilleau et al. 2022; Duran et al.
61 2022), as well as the detection of the martian transition zone (Huang et al. 2022) and core (Stähler
62 et al., 2021; Khan et al., 2022). Recently, the detection of surface waves enabled the
63 characterization of crustal structure variations away from the *InSight* landing site (Kim et al.,
64 2022).

65 The first observed Rayleigh waves originated from two large meteoroid impacts, called
66 S1000a and S1094b (Posiolova et al., 2022; see also Horleston et al., 2022 & Ceylan et al., 2022).
67 The photographic identification of the craters associated with the two events simplified the
68 interpretation of surface wave dispersion measurements by fixing the hypocentral depth and
69 location (Kim et al., 2022). However, the limited frequency content and the absence of Love waves
70 in the record prevented crustal structure below 30 km and anisotropy associated with the martian

71 crust to be constrained. In addition, the paths of the minor-arc Rayleigh waves (R1) were mostly
72 confined to the crust in the northern lowlands in the vicinity of Elysium (Kim et al., 2022). A
73 weaker major-arc Rayleigh wave (R2) arrival was also reported, but due to unclear polarization of
74 the data, the interpretation of structural constraints provided by this seismic phase was limited.

75 The marsquake S1222a, the largest event recorded during the mission to date (Mw 4.7;
76 Kawamura et al., 2022), provides an important opportunity to further constrain lateral variations
77 in crustal structure using surface wave analysis. Based on the epicentral location estimated by the
78 Marsquake Service (MQS; InSight Marsquake Service, 2022), the surface waves identified in
79 S1222a travel along the dichotomy on Mars (Figure 1A). Compared to the recordings of the two
80 large impacts, the identified surface waves in S1222a have a broader frequency content and include
81 not only Rayleigh but also Love waves, overtones, and multiple-orbit surface waves (e.g., Panning
82 et al., 2022). The expected sensitivity of the dispersion measurements extends down to the upper
83 mantle. Fortunately, the propagation paths for the surface waves are close to the dichotomy
84 boundary, where we expect crustal thickness variation to be greatest (Wieczorek et al., 2022)
85 (Figure 1A).

86 Here, we report robust group velocity measurements of Rayleigh and Love waves and their
87 first overtones for event S1222a. Using the available frequency content from the surface waves,
88 we invert the group velocities to obtain profiles of S-wave velocity and radial anisotropy down to
89 ~90 and 50 km depth, respectively. We compare our results with previously published models
90 derived from group velocity dispersion of fundamental-mode Rayleigh waves based on S1000a
91 and S1094b (Kim et al., 2022) and discuss implications for lateral variations in crustal structure
92 across the topographic dichotomy on Mars.

93 **2 Data and Methods**

94 **2.1 Surface wave dispersion measurements**

95 We use the seismic recording of S1222a and remove glitches from the 20 sample per second UVW
96 channels of the Very Broad Band (VBB) sensor (Scholz et al., 2020). The deglitched records are
97 rotated to ZNE and we confirm that the seismic waveforms are not strongly affected by any known
98 electro-mechanical noise by the sensor or the lander (Ceylan et al., 2021; Kim et al., 2021b). The
99 P- and S-arrivals are prominent both in the time and spectral domains across the 1-10 s period
100 range and have been assigned a picking uncertainty of ± 0.5 s and ± 2 s by MQS, respectively. Using
101 the differential travel time and P-wave polarization estimates from these body waves, MQS located
102 this event at a distance of $37 \pm 1.6^\circ$ and a back azimuth range between 98° and 121° from the lander
103 near Cerberus Plains (Figure 1). About 200 s after the S-arrival, strong dispersive arrivals are
104 evident in vertical- and horizontal-component data and are identified as minor-arc surface waves
105 by the MQS (InSight Marsquake Service, 2022). In contrast to previously reported surface waves
106 in S1094b (Kim et al., 2022), the S1222a recording shows both minor-arc Rayleigh (R1) and Love
107 waves (G1) on the vertical and transverse components, respectively. Assuming that propagation
108 occurs along the great circle path (GCP), the apparent time delay between R1 and G1 suggests an
109 anisotropic structure on Mars. Overtones and multiple orbit surface waves have also been detected
110 and cataloged by the MQS. See Kawamura et al., (2022) for more detailed information on event
111 description.

112 To make group velocity dispersion measurements on the identified surface wave arrivals,
113 we employ a single-station approach using a multi-wavelet transformation as a filter bank

114 (Poppeliers & Preston, 2019). Because the wavelet transform optimizes the trade-off between time
115 and frequency resolution compared to typical narrow-band filtering, this method achieves stable,
116 high-resolution measurements, while providing robust error estimates (Preston et al. 2020). Here,
117 we focus on a 400 s-length window around the surface wave arrivals and use 10 mutually
118 orthogonal wavelets to compute 10 dispersion estimates across the 5-50 s period range of the
119 vertical- and horizontal-component waveforms. Long-period energy beyond 50 s is visible but due
120 to presence of strong atmospheric noise, we focus on periods <50 s that have higher signal-to-
121 noise ratios (SNR) (Kawamura et al., 2022). For each period, we normalize to unity the power of
122 the resulting transform, and pick the maximum envelope amplitude for each of the 10 transforms
123 across different periods (e.g., Figure 2). Picks on vertical and radial components are collected for
124 Rayleigh waves while those on the transverse component are used for Love waves. Next, these
125 initial picks are filtered based on the back azimuth and polarization analysis described below.

126 2.2 Back azimuth and polarization analysis

127 To obtain robust dispersion measurements for inversion, we implement two additional steps that
128 discard those measurements that substantially deviate from the propagation direction and exhibit
129 particle motion inconsistent with that expected for surface waves. For Rayleigh waves, we perform
130 a grid-search to find back azimuth estimates that maximize correlation between the radial-
131 component and the Hilbert transform of the vertical component (i.e. maximize elliptical particle
132 motion in the vertical plane; see Figure 2A). For Love waves, We apply a similar grid-search to
133 the analysis window around G1 and its overtone as we minimize the ratio between the average
134 power of vertical and transverse component data. We find consistent back azimuth estimates
135 between body and surface waves with a small offset of $\sim 10^\circ$ possibly indicating complexities
136 associated with lateral varying structure along the wave propagation paths. However, we do not
137 have sufficient sensitivity to resolve such a small back azimuth difference observed in the data.

138 Next, we conduct frequency-dependent polarization analysis (Park et al. 1987) on the
139 S1222a waveforms to investigate the particle motion of the surface wave arrivals. We employ the
140 S-transform (Stockwell et al., 1996) of the three-component event waveforms and compute a 3x3
141 cross-component covariance matrix at each frequency in 90% overlapping time windows whose
142 duration varies inversely with frequency. The relative sizes of the eigenvalues of this covariance
143 matrix are related to the degree of polarization of the particle motion, while the complex-valued
144 components of the eigenvectors describe the particle motion ellipsoid in each time-frequency
145 window. Our computed polarization attributes (see Table S2-1 of Stähler et al. (2021)) are then
146 combined into a metric which highlights signals with elliptically-polarized energy in the vertical
147 plane for Rayleigh waves (e.g., Kim et al., 2022; Figure 2B). For Love waves, we examine the
148 phase angle of the particle motion ellipse to ensure our picks have particle motion that is
149 dominantly polarized in the horizontal plane. We discarded picks that show deviations away from
150 the expected propagation direction larger than the measurement uncertainty (Kawamura et al.,
151 2022) or have irregular polarization. The remaining picks resulting from both back azimuth and
152 polarization analyses are considered as our final measurements for inversion (Figure 2C). This
153 includes four dispersion curves for R1, G1, and their corresponding overtones in the 8-40 s period
154 range. Unlike R1, the direction of propagation and particle motions of the identified R2 and R3
155 are largely scattered and unclear due to low SNR. Based on our analysis, only a few measurements
156 of R2 and R3 are available at ~ 35 s thus we do not use the suggestive R2 and R3 arrivals directly
157 in the inversion.

158 2.3 Inversion of surface wave dispersion data

159 We invert the group velocity measurements summarized in Figure. 2C using a Markov chain
 160 Monte Carlo (MCMC) method for sampling the path-averaged S-wave velocity structure with an
 161 adaptation of the Metropolis-Hasting algorithm (Hastings & Keith, 1970). We assume a fixed
 162 V_P/V_S ratio of 1.81 estimated for the upper crust beneath *InSight* using a free surface transform
 163 matrix (Kim et al., 2021a). The scaling between V_P and density is based on Birch's law (Birch,
 164 1961). We employ a fixed parameterization strategy using b-splines as described in MCMC
 165 Approach 1 of Kim et al. (2022). We parameterize the crust using eight b-spline functions
 166 overlying a mantle halfspace with a constant velocity. The depth to this constant-velocity layer is
 167 allowed to vary between 30 – 70 km, the depth range estimated for the average crustal thickness
 168 of Mars (Wieczorek et al., 2022). We consider uniform *prior* distributions for those spline
 169 coefficients for V_S and radial anisotropy, i.e., $(V_{SH}-V_{SV})/V_S$ in the inversion, but present the
 170 anisotropy as $\xi = (V_{SH}/V_{SV})^2$ for easier comparison to other studies. Positive anisotropy
 171 corresponds to $\xi > 1$. Highest-accuracy computation of dispersion in a transversely isotropic
 172 medium would require the specification of 5 elastic parameters; here, we assume $V_{PV} = V_{PH}$,
 173 anellipticity or $\eta = 1$, and use V_{SV} and V_{SH} to compute the dispersion for Rayleigh and Love waves,
 174 respectively. This choice dramatically improves computational efficiency, while having negligible
 175 impact on accuracy in the period range used (Jiang et al., 2018; see also Beghein et al., 2022 for
 176 comparison of different parameterization schemes employed for anisotropic inversion). We
 177 compute chi-squared misfit between predicted and observed group velocity dispersion curves
 178 assuming a measurement uncertainty of 0.1 km/s based on the MQS event uncertainty (Kawamura
 179 et al., 2022). We vary both the assumed *a priori* distribution and the total number of MCMC
 180 iterations to ensure that our final model is not biased by these choices. The Rayleigh and Love
 181 wave group velocity kernels calculated using the mean posterior velocity model yield the
 182 sensitivity of available group velocity measurements is weak at depths shallower than 5 km, similar
 183 to the case with S1000a and S1094b (Kim et al. 2022). However, the longest-period R1 and
 184 overtones in S1222a extend the previously-reported sensitivity of ~5 – 30 km to ~90 km depth
 185 (~50 km depth for G1) (Figure 3A-B).

186 **3 Results and Discussion**

187 Our V_{SV} profile is characterized by a positive velocity gradient of 0.015 km/s per km with an
 188 average velocity of 3.45 km/s between ~5-60 km depth (Figure 3C). Like models derived from
 189 Rayleigh wave group velocity measurements from the two large impacts (S1000a & S1094b),
 190 which have the average V_{SV} of 3.2 km/s between ~5-30 km depth (Kim et al., 2022), the S1222a
 191 models also show little depth-variation in V_{SV} with a slightly slower average V_{SV} of 3.17 km/s in
 192 the overlapping sensitivity depth ranges. Below 30 km depth, the posterior distribution of S1222a
 193 V_{SV} is shown to be compatible with the broader distribution of S1000a V_{SV} (cyan, Fig. 3C),
 194 approaching 3.8 km/s at ~60 km. We observe a large discontinuity in the V_{SV} profile at 63 ± 10
 195 km depth with a velocity jump of ~0.4 km/s (representing a total impedance contrast of ~20%).
 196 This velocity jump accounts for the steep increase in the group velocity of R1 and its overtone
 197 seen near 25 s and 12 s, respectively. Similar velocity increases at periods with sensitivity near the
 198 crust-mantle boundary are observed on Earth in both continental and oceanic settings (e.g., Ewing
 199 and Press, 1950; Ewing and Press, 1952). Below the discontinuity, the average velocity of 4.1 km/s
 200 is consistent with that inferred for the upper mantle for Mars from body wave analyses (Khan et
 201 al., 2021; Duran et al., 2022). Therefore, we interpret the ~63 km deep interface to represent the
 202 crust-mantle boundary along the R1 path. Its depth is well within global crustal thickness estimates

203 on Mars (Wieczorek et al., 2022). The abruptness of the velocity jump across the discontinuity is
204 in part due to how the mantle property is being parameterized by a constant value. Moreover,
205 because the surface wave sensitivity functions are fundamentally broad over a wide depth range
206 (e.g., Figure 3A-B) and the R1 path traverses near the dichotomy, we are unable to constrain
207 whether the martian crust-mantle boundary is sharp or gradational (Figure S4).

208 Identification of both R1 and G1 in S1222a allows us to determine radial anisotropy of
209 shear wave speeds within the crust. We find that a model where radial anisotropy steadily decreases
210 from $\xi \sim 1.3$ (equivalent to 12% for $(V_{SH} - V_{SV})/V_S$) at ~ 5 km depth with a gradient of -0.01 per
211 km depth is required to fit the dispersion measurements of the Love waves (Figure 3D). Our
212 resolution test on the anisotropic inversion shows that the observed anisotropy in the model is
213 resolvable within the sensitivity depth range of the surface wave data (Figure S4). Indeed, no
214 simple isotropic crust can explain the group velocities of G1 or its first overtone if we preclude the
215 presence of anisotropy in our inversions (Figure S5). While our posterior distribution of ξ prefers
216 values less than one ($V_{SV} > V_{SH}$) in the lower crust (below 40 km) we do not believe this to be
217 robust because we lose sensitivity below ~ 50 km depth due to the absence of long-period Love
218 wave dispersion measurements; this is reflected in the large model uncertainties between ~ 50 -60
219 km depth. Furthermore, we observe little change in the Chi-squared misfit when values for $\xi > 1$
220 are allowed during the inversion (Fig. S6 and S9). Variation of data uncertainty does not strongly
221 affect our inversion results either (Fig. S7-S9).

222 To quantify the implication of our inverted models, including the observed crustal
223 thickness in a global context, we generate Rayleigh wave phase / group velocity maps of Mars by
224 linearly extrapolating our S-wave velocity profile based on the crustal thickness model constrained
225 by gravity data (Wieczorek, 2021). Following the modeling steps described in Wieczorek et al.
226 (2022), each crustal thickness model is produced by fixing the crustal thickness to 45 km at the
227 location of the lander (Knapmeyer-Endrun et al., 2021) assuming a density contrast over the
228 boundary of topographic dichotomy as mapped by Andrews-Hanna et al. (2008). We compute
229 global crustal thickness models for spherical harmonics up to degree 120. Then, we linearly scale
230 the velocity profiles in Fig. 3C and from Kim et al. (2022) with the relative depth variations to the
231 crust-mantle boundary. Motivated by the crustal models of Mars discussed in Wieczorek et al.
232 (2022), two end-member dichotomy models are tested: (i) type 1 - a model with a uniform density
233 ranging from 2550 kg/m^3 to 3050 kg/m^3 (e.g., Baratoux et al., 2014; Wieczorek et al., 2019) and
234 (ii) type II - a model with a density contrast of 100 - 500 kg/m^3 across the dichotomy that may
235 indicate crust that originated exogenically (e.g., Andrews-Hanna et al., 2008; Nimmo et al., 2008)
236 (Figure 4A-B).

237 Due to a known trade-off between crustal thickness and density, the larger the assumed
238 density contrast, the smaller the average crustal thickness variation across the dichotomy becomes.
239 Hence, these models represent two limiting cases for describing the crustal dichotomy structure on
240 Mars. We compute predicted travel times of the R1 as well as R2 and R3 arrivals by kinematic ray
241 tracing of the surface waves through phase velocity maps at different periods (e.g., equations
242 16.185 & 16.186 in Dahlen and Tromp, 1998). We find that the resulting travel times deviate by
243 less than 1% between the GCP and the ray theoretical path for R1-R3 arrivals (Fig. S10), justifying
244 our use of the great circle approximation in the inversions. Therefore, lateral thickness variations
245 are unlikely to substantially affect the travel times of the surface waves particularly for long-
246 periods > 25 s. However this assumption may not be optimal for higher-orbit surface waves beyond
247 R3 (Fig. S10) and a 3D sensitivity kernel should be taken into account rather than a simple range
248 of GCPs drawn from the back azimuth uncertainties. More realistic 3D wavefield simulations

249 through candidate crustal thickness models are beyond the scope of this study, but such an effort
250 should provide further information on 3D surface wave propagation (e.g., Bozdog et al., 2017).

251 In the group velocity maps shown in Figure 4C-D, we gray out regions outside the vicinity
252 of the GCP based on the back azimuth uncertainty of the event location. The R2 and R3 arrivals
253 in our S1222a data travel with an average speed of 2.88 km/s traversing much greater distances
254 along the potential velocity contrast across the dichotomy. This speed is 0.13 km/s faster than the
255 R1 at ~ 35 s period (Figure 2C). We find the spread in group velocities due to GCP propagation is
256 substantially smaller than the range predicted by different crustal thickness models (Figure 4E-H).
257 For example, the type II model with a 300 kg/m^3 density contrast at 35 s implies an average crustal
258 thickness of ~ 43 km which yields $\sim 7\%$ larger average group velocity for R2 than the uniform
259 density model. Because the crust in type I model is thicker on average, the corresponding velocity
260 profile is stretched downward to larger depths and the apparent speed at which the surface waves
261 travel at a given depth is expected to be slower for such a type of model. The group velocity
262 predictions for R2 and R3 arrivals in S1222a are strongly dependent on the choice of velocity
263 profiles used in the modeling (Figure 4E-H). While type I crustal model shows the best-fit between
264 the group velocity predictions and measurements of R2 and R3 arrivals based on the previous
265 velocity profile of S1094b, we were not able to explain the data with the new V_{SV} profile of S1222a
266 (Figure 3C) and the predictions are largely under-estimated. If we assume the scaling approach in
267 the global extrapolation is optimal, our analysis suggests that the low velocity structure beneath
268 the lander (Knapmeyer-Endrun et al., 2021) which is also evident below the S1222a R1 path
269 (Figure 3C) may not be prevalent along the equatorial dichotomy. Crucially, the average Moho
270 depth of ~ 63 km shown in our inversion is only compatible with the absence of a density contrast
271 across the dichotomy (inset, Figure 4C-D), independent of the velocity profiles used in the
272 extrapolation procedure (Fig. S11-S22).

273 Regardless of the choice of models, the Hellas impact basin is expected to be a significant
274 outlier, with velocities close to that of the mantle across different periods (Figure 4). Once
275 corrected for the fraction of the path traversing Hellas, it has been suggested that the crustal wave
276 speed at 5-30 km depth is similar between the northern lowlands and the southern highlands (Kim
277 et al. 2022). Here, we provide another independent constraint indicating that there is no large-scale
278 dichotomy in average crustal density. We also constrain the crustal velocity structure at those
279 depths to be largely similar (difference less than 5%) with a caveat that the non-linearity of the
280 surface wave sensitivity to depth is difficult to be implemented with such sparse data collected on
281 Mars. At greater depths, the propagation path samples both with crust and mantle, and periods
282 larger than 50 s using the higher-order multiple orbit surface waves would have to be further
283 analyzed and reviewed (InSight Marsquake Service, 2022; Kawamura et al., 2022).

284

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295 **Open Research**

296 The InSight event catalogue (<https://doi.org/10.12686/a17>) and waveform data are available from
297 the IRIS-DMC (<http://ds.iris.edu/ds/nodes/dmc/tools/mars-events/>), NASA-PDS ([https://pds-
298 geosciences.wustl.edu/missions/insight/seis.htm](https://pds-geosciences.wustl.edu/missions/insight/seis.htm)) and IPGP data center
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300 produced by resources available at <https://doi.org/10.5281/zenodo.4439426>.

301

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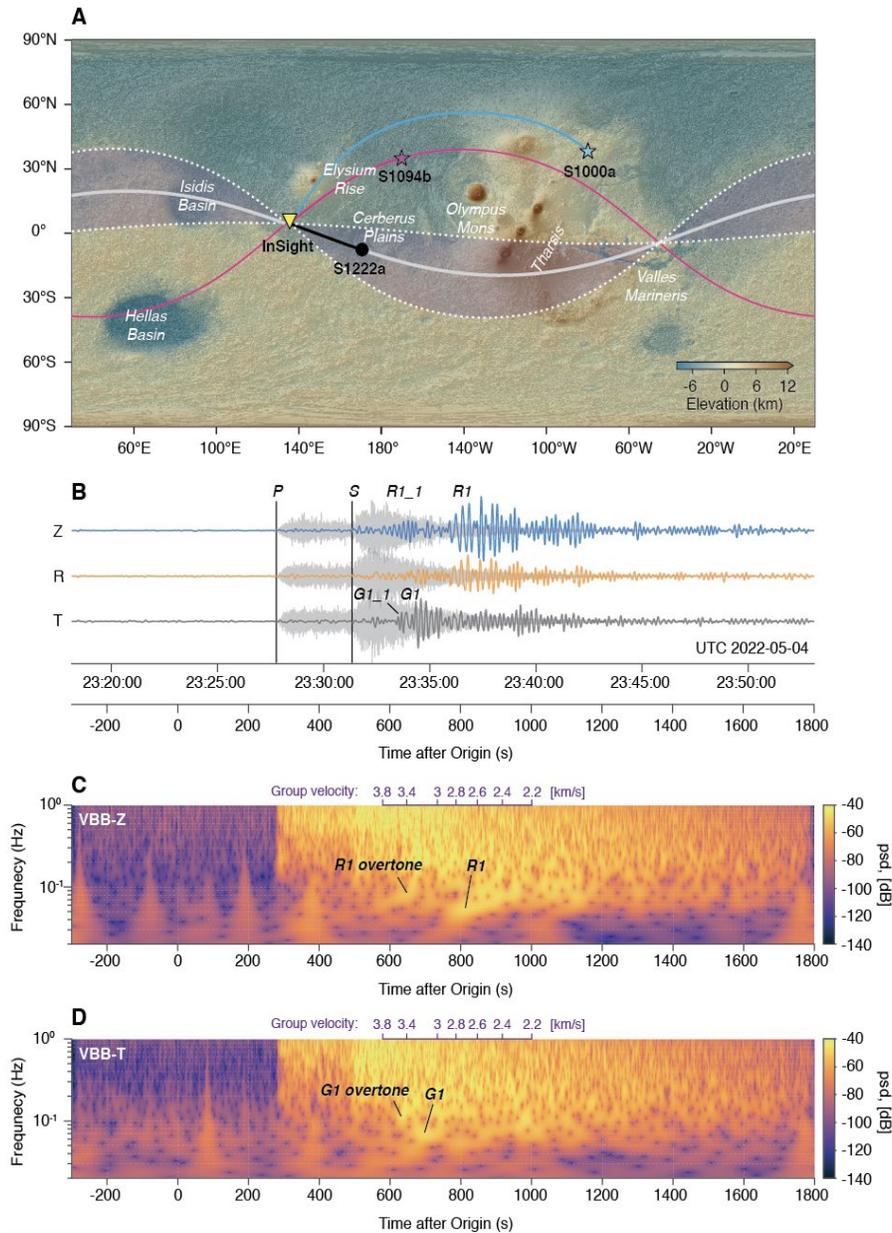
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419 **Figures & Figure captions:**

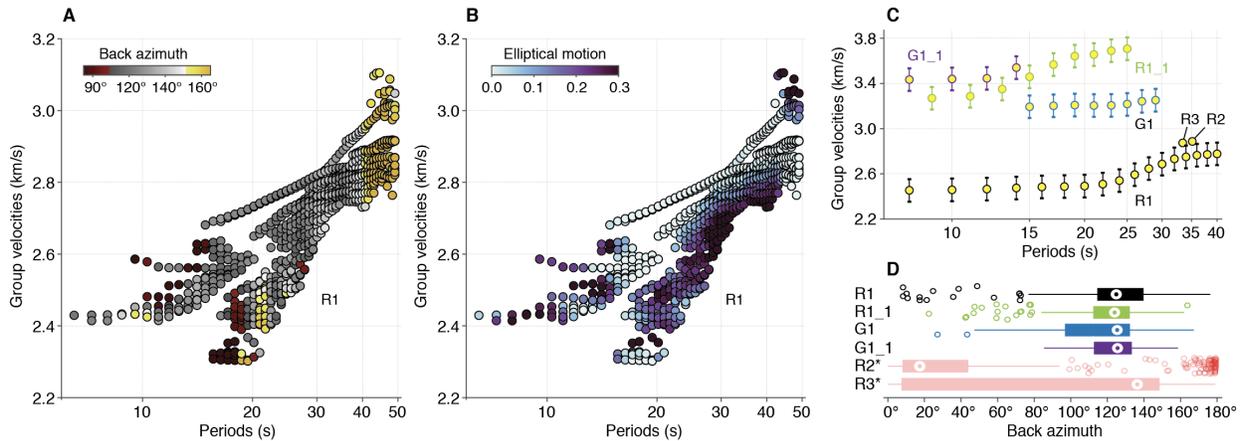
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422 **Figure 1.** (A) Location of the event S1222a (black symbol). The lander location is denoted by the
 423 yellow symbol. The great circle paths for minor-arc (R1, black) and major-arc Rayleigh waves
 424 (R2, gray) in S1222a are in solid, while those paths including the back azimuth uncertainty are
 425 displayed in dashed line. Location of two large meteoroid impacts (S1094b and S1000a) and the
 426 corresponding paths for previously identified surface waves are based on Posiolova et al. (2022)
 427 and Kim et al. (2022), respectively. (B) Broadband three-component (ZRT) seismogram of S1222a
 428 (light gray) with P and S wave picks. Rayleigh and Love waves are clearly visible on the data
 429 bandpass filtered between 10-100 s. R1_1 and G1_1 denote the first overtones of R1 and G1,
 430 respectively. (C) Vertical and (D) transverse component S-transforms show large amplitude
 431 surface wave arrivals with dispersion. Time after origin of the event is converted to group velocity
 432 using the equatorial radius of Mars (purple ticks with labels at the top of each spectrogram).
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436 **Figure 2.** (A) Back azimuth and (B) frequency dependent polarization estimates for group velocity
 437 measurements of the R1 arrival. Measurements from both vertical and radial component in panels
 438 2A and 2B are based on a multi-wavelet approach. The corresponding results including G1, R2,
 439 R3, and the overtones of R1 and G1 are in the supporting information Figure S1-3). (C) Summary
 440 of Rayleigh and Love wave dispersion measurements. R1_1 and G1_1 denote the first overtones
 441 of R1 and G1, respectively. Measurements from the off-great-circle propagation with low elliptical
 442 polarization (less than 0.1 above the global average) are discarded. Uncertainty of 0.1 km/s is
 443 assigned to our measurements to account for the MQS epicentral uncertainty of the event. (D)
 444 Distribution of the back azimuth estimates for the surface wave arrivals analyzed in this study. The
 445 maximum whisker length is specified as 1.0 times the interquartile range. Outliers beyond the
 446 whiskers are denoted by circle symbols. Note, the back azimuth estimates for the R2 and R3
 447 arrivals are more scattered compared to those from the minor-arc surface waves due to low SNR
 448 of the data.

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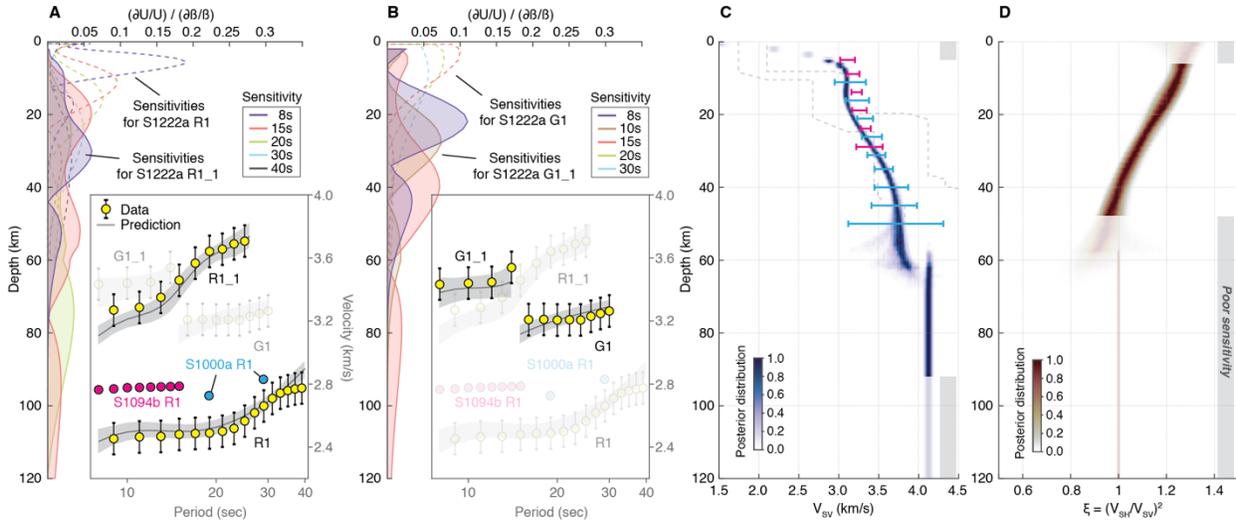
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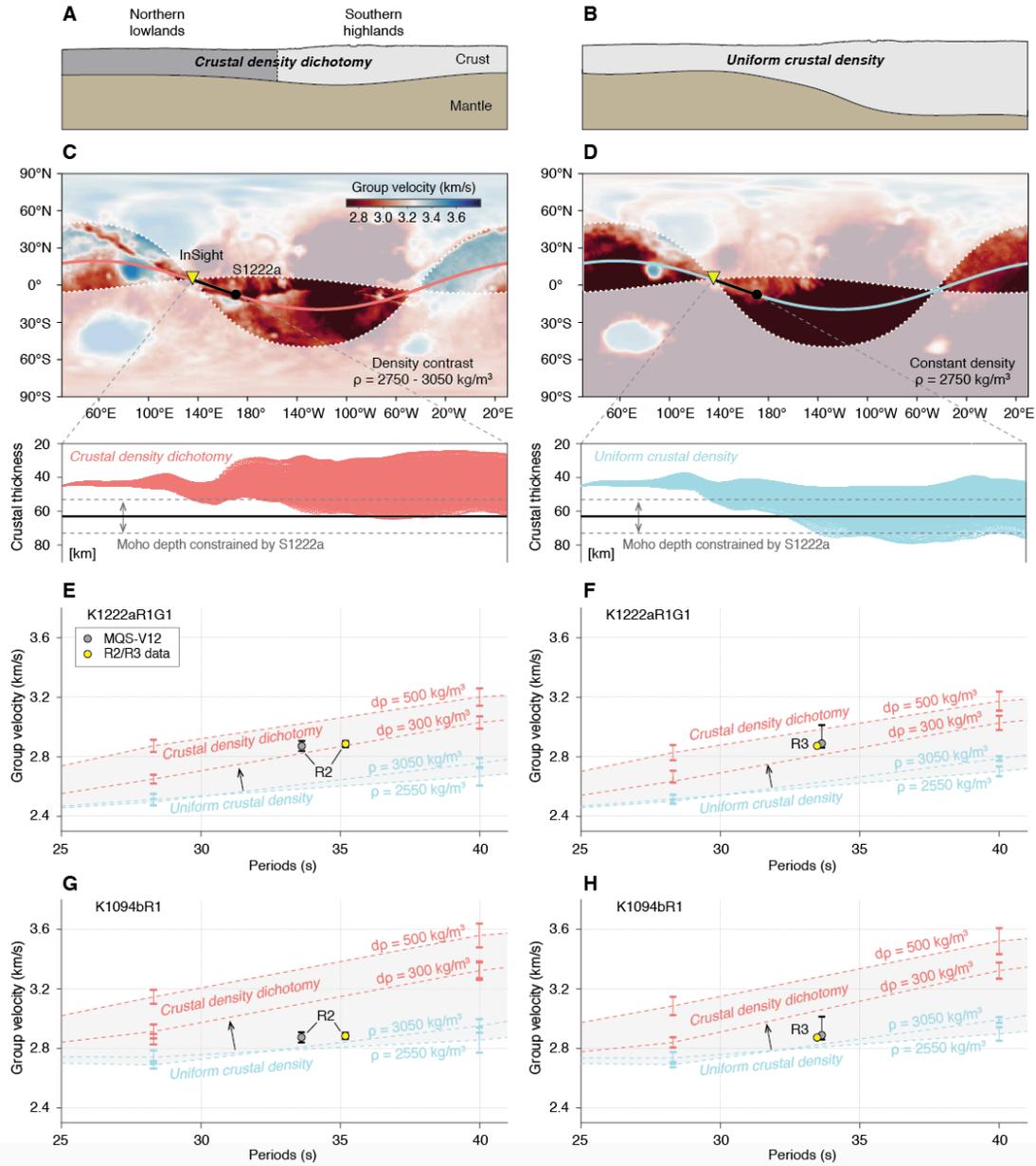
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Figure 3. (A-B) Depth sensitivity kernels and data vs. prediction (inset) of the fundamental-mode (dashed lines) Rayleigh waves, Love waves, and the overtones (solid lines with enclosed area shaded). Kernels are computed based on the average model from the inversion. R1_1 and G1_1 denote the first overtones of R1 and G1, respectively. Mean predicted dispersion curves are denoted by gray lines. (C) Posterior distribution of V_{sv} and (D) radial anisotropy structure inverted from the group velocity dispersion curves of S1222a. Posterior distribution and prediction are based on the best-fitting 10,000 models after two million iterations. Depths where sensitivity is inadequate (<40% in cumulative kernel strength) are muted. Note our V_{sv} is constrained by a combination of both Rayleigh and Love waves while the radial anisotropy ξ is primarily constrained by Love waves. Posterior distributions from the isotropic inversion of S1000a (light blue) and S1094b R1 (magenta) are denoted by horizontal lines at each depth (Kim et al., 2022).



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471 **Figure 4.** Conceptual models of the martian crust (A) with and (B) without a density contrast and
 472 the corresponding global group velocity maps at 40 s using the inverted velocity profile of surface
 473 waves in S1222a (C-D), respectively. Regions shaded indicate the off-propagation paths based on
 474 the back azimuth analysis. Inset below (C-D) shows crustal thickness profiles along the S1222a
 475 R1 for each type of models for all tested density values in Fig. S11-S16. Horizontal solid and
 476 dashed lines indicate the average crustal thickness and its uncertainty observed in Fig. 3C. (E-F)
 477 Average group velocities along the propagation paths for all of the models using the velocity
 478 profile in S1222a (K1222aR1G1; Fig. S11-16) and (G-H) S1094b (K1094bR1; Fig. S17-S22).
 479 Group velocity at each period increases as density contrast increases across the dichotomy
 480 boundary. Note the uncertainties of the R2 and R3 arrivals in the MQS catalog V12 (gray symbol)
 481 are substantially larger than those provided by this study (yellow symbol) due to the absence of
 482 back azimuth and polarization constraints.