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Crustal Anisotropy in the Martian Lowlands from Surface Waves

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Key Points:

- Rayleigh and Love waves were detected on Mars;
- Seismic wave speed is directionally dependent over large-scales within Mars crust;
- The shear wave horizontal velocity is faster than the vertical velocity at 10 – 25 km depth in the lowlands between the event and the lander.

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Abstract

The largest seismic event ever recorded on Mars, with a moment magnitude of 4.7 ± 0.2 , is the first event to produce both Love and Rayleigh wave signals. We measured their group velocity dispersion between about 15 s and 40 s period and found that no isotropic depth-dependent velocity model could explain the two types of waves simultaneously, likely indicating the presence of seismic anisotropy. Inversions of Love and Rayleigh waves yielded velocity models with horizontally polarized shear waves traveling faster than vertically polarized shear waves in the top 10 km – 25 km. We discuss the possible origins of this signal, including the preferred orientation of anisotropic crystals due to shear deformation, alignment of cracks, layered intrusions due to an impact, horizontal layering due to the presence of a large-scale sediment layer on top of the crust, and alternation of sedimentation and basalt layers deposits due to large volcanic eruptions.

Plain Language Summary

The largest marsquake recorded so far sent energy across the red planet in the form of different kinds of waves, including two types of waves trapped near the surface. This was the first time both types of so-called surface waves were detected on Mars. Combining measurements from these two types of surface waves allowed us to determine the speed of other types of waves, i.e. waves that travel horizontally and that make rocks move perpendicular to the direction of propagation. We found that these waves move faster in the crust between 10 km and 25 km depth when the rocks oscillate in a direction sub-parallel to the planet surface than if the rocks vibrate in the vertical direction. This wave speed dependence can tell us about deformation mechanisms inside the crust. We found that either an alternation of volcanic rocks and sediments layers due to volcanic eruptions or internal layering within the crust due to an impact are the preferred explanations for our observations.

1 Introduction

The largest seismic event ever detected on Mars occurred on May 4 2022, and was recorded by the Seismic Experiment for Interior Structure (SEIS) Very Broadband (VBB) seismometer (Lognonné et al., 2020) of NASA’s InSight mission (Banerdt et al., 2020). The event, labeled S1222a, occurred at 23:23:06.57 UTC (sol 1222) and has a moment

magnitude M_w estimated at 4.7 ± 0.2 (Kawamura et al., 2022). This event is thus five times larger than the second largest event recorded (Horleston et al., 2022) and had enough energy to allow for the first direct observation of multi-orbiting Rayleigh waves on Mars (Panning et al., 2022). Based on their calculated back azimuth of 101° (with a range of $96^\circ - 112^\circ$) and epicentral distance ($37^\circ \pm 1.6^\circ$), the Mars Quake Service (MQS) determined that this quality A (InSight Marsquake Service, 2020) event originated at 7.63° S, 170.67° E, near the North-South dichotomy, east of the landing site, and south of Cerberus Fossae (Fig. 1).

Most remarkably, S1222a generated surface waves clearly observable on all three components of the ground motion (Fig. 2). While Rayleigh waves have been previously detected on Mars with impact events S1000a and S1094b (Kim et al., 2022), no Love waves had been seen before. Since Rayleigh and Love waves are predominantly sensitive to different elastic parameters (governing the speed of vertically and horizontally polarized shear waves travelling horizontally, respectively), this gives us an opportunity to study the presence of large-scale radial seismic anisotropy on Mars between the lander and the event epicenter. Radial anisotropy, which is a type of transverse isotropy with a vertical symmetry axis, manifests as a difference in the velocities of vertically (V_{SV}) and horizontally (V_{SH}) polarized shear waves. It is quantified by parameter $\xi = V_{SH}^2/V_{SV}^2$ and is a powerful tool to study the deformation history of a planet (e.g. Weiss et al. (1999)).

2 Data Analysis

We first removed the instrument response from the raw data to get the velocity record, and rotated the U, V, and W components of the SEIS instrument to N, E, and Z. We then rotated the N and E components to the radial (R) and tangential (T) coordinates using a back azimuth of 101° as determined by the MQS (Kawamura et al., 2022). The waveforms display clear wave packets on the Z and R components, characteristic of Rayleigh waves, and a wave packet on the T component, identified as a Love wave (Fig. 2(a)). Using two-pass, second-order Butterworth filters in different frequency bands (10 – 20 s, 15 – 25 s, 20 – 30 s, 25 – 30 s, and 30 – 40 s), we found that S1222a generated surface wave energy at periods between about 10 s and 40 s on all components of the seismogram (Fig. S1).

We subsequently carried out a multiple filter analysis (MFT) of the Z, R, and T components to isolate the group velocities as a function of time (Dziewonski et al., 1969; Herrmann, 2013). This consists in narrow bandpass filters using the operator $\exp[-\alpha(\omega - \omega_0)^2/\omega_0^2]$ where ω_0 is the center frequency. Parameter α compromises between resolution in the time and frequency domains. We measured the group velocities by picking the maximum energy of the waveform envelope for each period (Fig. 2(b) – (d)). The velocity uncertainty is defined by a threshold energy of at least 1 dB of the energy maxima as in Beghein et al. (2010). The MFT analysis shows that Rayleigh waves have peak energy between about 15 s and 40 s period and group velocities increasing continuously up to about ~ 3 km/s (Fig. 2(b) – (c)). The transverse component displays peak Love wave energy in the same period range but with faster group velocities between about 3.2 km/s and 3.5 km/s (Fig. 2(d)). We note that the T component MFT plot differs from the Z and R components in that its dispersion curve first decreases between 10 s and 15 s and becomes relatively flat at greater periods.

We tested that the effect of back azimuth uncertainties on the dispersion curves and found no significant change (Fig. S2). We also performed an ellipticity analysis in different period bands (Fig. S3). The elliptical and retrograde motion found between 15 s and 35 s confirm the Rayleigh wave nature of the signal.

We additionally observe discrepancies in the Rayleigh wave group velocities between the Z and R components at periods >30 s, and even more so at periods >35 s for which there is little overlap between the error bars (Fig. 2 and Fig. S3(a)). These differences could be due to the lower waveform amplitude along the R component (Fig. S3(b)), making it more easily contaminated by noise. We considered stacking the R and Hilbert-transformed Z components to improve the signal-to-noise ratio and obtain more reliable and consistent group velocities at periods >30 s. However, our ellipticity analysis showed that, while at short period (15 – 30 s) the phase shift between the Z and R components is about 90° (i.e., the major and minor axes of the ellipse are not parallel to the x- or y-axis), it is not the case at 30 – 35 s, indicating that applying a Hilbert transform will not help. At periods of 35 – 40 s, although the phase shift is close to 90° , the ellipticity (i.e., the ratio between Z and R) is much larger than that at short periods due to smaller amplitudes on the R component and may we not have a reliable signal. Because of these discrepancies between the Z and R components, we chose to invert measurements up to 30 s

only. Inversions including all periods were performed for comparison. They are presented in the Supplementary Material and discussed in section 4.

The presence of transient signals, including glitches due to diurnal temperature variations on Mars, has the potential to contaminate the waveforms (e.g. Kim, Davis, et al. (2021)). We performed a glitch detection on the raw data using a synthetic glitch template (Scholz et al., 2020) and found no glitches near the first minor-arc surface wave arrivals (Fig. S4).

3 Method

The measured Love and Rayleigh wave dispersion curves were inverted jointly to obtain one-dimensional (1-D) V_{SV} and V_{SH} models representing the average structure between the station and the quake. The V_{SH} and V_{SV} model distributions were then combined to calculate radial anisotropy ξ . We used the Computer Programs in Seismology (Herrmann, 2013) code package to generate the synthetic group velocity, and a Niche Genetic Algorithm for the inversion (Koper et al., 1999; Li et al., 2021) together with a B-spline parameterization, as described below. The results obtained with this approach, hereafter referred to as Method 1, were compared against results from two other techniques (Method 2 and Method 3). Method 2 is a fully non-linear transdimensional hierarchical Bayesian approach using a reversible jump Monte Carlo Markov Chain (MCMC) algorithm and a layered parameterization that is allowed to vary at each iteration (Weidner et al., 2022). Method 3 also uses a MCMC approach but the parameterization is fixed and uses B-splines. These methods are described in more details in the Supplementary Material.

The period range at which the detected surface waves display strong energy (15 – 30 s) corresponds to sensitivity to structure in the upper $\sim 70 - 80$ km, though it is important to note that Love wave sensitivity to V_{SH} is greatly reduced below 50 km (Fig. S5). Anisotropy was thus allowed in the models down to 80 km depth.

In the 0 – 80 km depth range, the V_{SV} and V_{SH} profiles were characterized by eight third-order B-splines. Since Rayleigh waves display some sensitivity down to about 130 km, we also inverted for isotropic shear wave velocities between 80 km and 130 km with a depth interval of 20 km for the depth anchor points. The prior constraint for all the shear wave velocities were uniform distributions from 1.5 km/s to 4.5 km/s. At depths greater than

130 km, we used the InSight model KKS21GP, which was extracted from the models of Khan et al. (2021) and Stähler et al. (2021) and selected to have a crust comparable to that of Knapmeyer-Endrun et al. (2021). The V_P/V_S ratio was set to be 1.80 and the density-to- V_S ratio was assumed to be 0.8. This choice was based on the ratios of the receiver function study of Knapmeyer-Endrun et al. (2021), which are between 1.7 to 1.9 for V_P/V_S and 0.7 to 0.9 for density-to- V_S . In total, we searched for 200,000 models during the inversion, and the acceptable models were selected if their predicted group velocity at each period fits the measurements within uncertainties. The posterior distributions are based on these best data fitting models.

4 Results

We first inverted the Rayleigh wave dispersion curve and used the resulting 1-D models to predict Love wave dispersion, and vice versa (Fig. S6). It is clear that isotropic models obtained with the Rayleigh wave data are unable to explain the measured Love wave dispersion and that the models obtained from Love waves alone cannot represent the Rayleigh wave measurements. This same behavior has been widely observed in Earth's upper mantle: Love waves usually have larger phase velocities than predicted by isotropic models that fit Rayleigh wave phase velocities. It has also been detected in the continental crust (e.g., Moschetti et al. (2010)). This apparent Love-Rayleigh discrepancy was first observed on Earth by Anderson (1961) and can be resolved by introducing transverse isotropy in the model (Harkrider & Anderson, 1962). We thus performed joint Love and Rayleigh wave inversions for V_{SH} and V_{SV} as described in section 3, and calculated the resulting anisotropy ξ and Voigt V_S average model distributions: $V_S = \sqrt{(2V_{SV}^2 + V_{SH}^2)/3}$ (Babuska & Cara, 2012).

Our average velocity models in the top 5 km and below 30 km depth exhibit significant uncertainties and variability between the different methods. The velocity obtained with Method 1 at shallow depths is around 3.0 ± 0.1 km/s. V_S then increases to 3.2 ± 0.06 km/s around 20 km depth and reaches 3.4 ± 0.1 km/s around 40 km below the surface. For Method 2, the V_S models are characterized by about two layers with $V_S = 3.2 \pm 0.1$ km/s in the top 20 – 30 km and $V_S = 3.4 \pm 0.2$ km/s below, down to 50 km depth. With Method 3, V_S increases from 2.5 ± 0.3 km/s to 3.0 ± 0.75 km/s within the top 5 km, $V_S = 3.4 \pm 0.8$ km/s at 20 km depth and $V_S = 3.5 \pm 0.2$ km/s at 40 km.

Our results also show a clear $V_{SH} > V_{SV}$ signal with $\xi \simeq 1.1$ between 5 km and ~ 30 km depth (Fig. 3), consistent across all three methods within uncertainties (Figs. S7 and S8). The posterior model uncertainties do not, however, rule out $V_{SH} < V_{SV}$ in the top 5 km and deeper than about 30 km. We additionally performed synthetic tests that confirmed these results: In Fig. S9, we created and inverted an input model with Method 1 that contained 10 % anisotropy in the 0 – 10 km depth range, 20 % at 10 – 30 km depth, and 10 % at 30 – 50 km. The model was isotropic at greater depths. The output revealed that, while the positive anisotropy is well recovered in the 10 – 25 km depth range, several solutions of opposite sign were found at shallower and greater depths. The lack of resolution in the top 5 – 10 km is likely due to the sensitivity of the data at the period range for which we have measurements. Shorter period data would be able to better constrain the shallower depths but have not been reliably recorded with the event analyzed here.

Fig. 3 corresponds to inversions including only the Rayleigh wave data that were consistent between the Z and R components, i.e. up to 30 s period. For completeness, we also inverted the data measured on the Z and R separately up to 40 s (Fig. S10). These tests show that including periods above 30 s yields model uncertainties that depend on which component the Rayleigh waves are measured. In addition, while most models still display $\xi > 1$ between 10 km and 25 km, the error bars are significantly larger. This dependence of the models on the data at periods greater than 30 s validates our decision to only present results based on the shorter period data in Fig. 3.

5 Discussion

5.1 Seismic Velocities

Lognonné et al. (2020) discussed the rock types, including sedimentary, impact-related, and volcanic, and their associated seismic velocities in the Martian crust around the In-Sight landing site. They found that the low velocities they had obtained for the shallowest crustal layer (upper ~ 8 km) suggest altered and/or damaged upper volcanic rocks. The velocities in the underlying layer (around 8 – 20 km depth), with $V_P = 4.5 - 6$ km/s and $V_S = 2.5 - 3.3$ km/s, were consistent with altered, iron-rich basaltic rocks. Similar velocities were later found by Knapmeyer-Endrun et al. (2021).

In our study, Methods 1 and 2 predict $V_S = 3.0 \pm 0.1$ km/s and $V_S = 3.2 \pm 0.1$ km/s, respectively, in the top 5 – 10 km, which is larger than the 1.7 – 2.1 km/s estimated by Lognonné et al. (2020) and Knapmeyer-Endrun et al. (2021) at the landing site. Method 3 gives lower V_S values between 2.5 ± 0.3 km/s and 3.0 ± 0.75 km/s in that depth range, but they are still larger than found beneath the lander. Using $V_P/V_S = 1.8$ (Knapmeyer-Endrun et al., 2021), we calculated that this corresponds to $V_P = 4.5 - 5.4$ km/s, which is still compatible with damaged or altered basalt, but also with impact melt, impact breccia, and sandstone (Lognonné et al., 2020).

Between ~ 10 km and 30 km depth, V_S is consistent among the three methods with values around 3.1 – 3.5 km/s. Using a $V_P/V_S = 1.8$ (Knapmeyer-Endrun et al., 2021), this correspond to $V_P \simeq 5.6 - 6.3$ km/s. Compared to Lognonné et al. (2020), our velocity models at those depths are thus compatible with volcanic rocks (damaged, altered, or compact basalt) and with impact melt.

5.2 Seismic Anisotropy

The presence of seismic anisotropy, i.e. the directional dependence of seismic wave velocity, is well documented on Earth (Dziewonski & Anderson, 1981; Montagner, 1994). In Earth’s upper mantle, it is considered a strong indicator of past (in the mantle lithosphere) or present (in the asthenosphere) strain coherent over large scales, resulting in textural changes such as lattice or crystal preferred orientation (LPO or CPO) of intrinsically anisotropic olivine crystals due to mantle deformation (Nishimura & Forsyth, 1989; Karato & Wu, 1993; Montagner, 1998; Becker et al., 2003; Yuan & Romanowicz, 2010).

In Earth’s crust, seismic anisotropy is often attributed to the shape preferred orientation (SPO) of isotropic material with contrasting elastic properties (Backus, 1962), such as the preferred alignment of melt pockets or lenses (Bastow et al., 2010), the distribution of fluid-filled cracks, or preferentially oriented pore spaces (Crampin, 1981; Crampin & Booth, 1985; Crampin, 1989). SPO can be caused by sedimentary layering, magmatic intrusions, or the presence of faults. In this case, the observed large-scale seismic anisotropy represents the effective medium comprised of unmapped fine (subwavelength thickness) isotropic layered structures (Backus, 1962). Alternatively, observations of anisotropy with $V_{SH} > V_{SV}$ in Earth’s mid-to-lower continental crust are often attributed to the alignment of minerals such as mica in ductile flow (e.g., Brownlee et al. (2017)). In Tibet, for

instance, it has been proposed that this type of anisotropy could be due to underthrusting of the Indian crust beneath Tibet or could result from ongoing crustal thinning (Shapiro et al., 2004).

Our results indicate a robust seismic anisotropy signal in the 10 – 30 km depth range (Fig. 3). Constraints from gravity data show that the crustal thickness along the event-station path, which samples the lowlands, is around 40 km (Wieczorek et al., 2022). In addition, recent P- and S-wave receiver function analyses with seismic data collected by SEIS have shown that the crustal thickness at the lander site is between 20 km and 40 km (Knapmeyer-Endrun et al., 2021), with a preference for 40 km with PP receiver functions (Kim, Lekić, et al., 2021).

In order to observe seismic anisotropy, there needs to be coherent preferred orientation over large distances. Specifically, producing $V_{SH} > V_{SV}$ would require either the preferential horizontal alignment of fast axes of anisotropic minerals or cracks. Therefore, we rule out the alignment of cracks (Crampin, 1981; Crampin & Booth, 1985; Crampin, 1989) as a likely explanation for the observed signal since alignment of vertically-oriented cracks would yield $V_{SV} > V_{SH}$. In addition, at depths greater than about 20 km, the pressure is such that cracks tend to close.

Another possibility is horizontal layering of isotropic material (with layer thicknesses much smaller than the dominant wavelength), i.e. extrinsic or apparent anisotropy. Where the different layers have contrasting elastic properties, an overall anisotropic medium results. In order to assess the degree of heterogeneity required, we have performed some simple calculations using Backus averaging (Backus, 1962). Fig. S11 shows the resulting ξ parameter for a range of degrees of shear wave velocity perturbations, and layer thickness ratios. Producing an anisotropic signal compatible with our observations requires a relative velocity perturbation of $\sim 20\%$, when fast and slow layer thicknesses are approximately equal. As either material begins to dominate the ensemble, the required perturbation significantly increases. Such a mechanism is observed in sedimentary basins on Earth (Sayers, 1998). On Mars, there is evidence for past floods and sedimentary transportation near the rim at the Pathfinder landing site (Rodriguez et al., 2019), as well as the presence of rivers and even an ocean that may have had enough water to fill the Vastitas Borealis, the Northern basin in the upper hemisphere (Carr & Head III, 2003; Di Achille & Hynek, 2010; Villanueva et al., 2015). Such a standing body of wa-

ter could have deposited sediments over large distances (Head et al., 2001). However, McLennan (2012) estimated the global average thickness of the sediment layer to be less than 2 km. Although the sedimentary rocks at several locations can reach greater depths, they are shallower than 10 km in many places (e.g., 5 km in Gale crater (Grotzinger et al., 2015), 3 – 6 km in Juventae Chasma (Grotzinger & Milliken, 2012), and 3 km in Medusae Fossae (Bradley et al., 2002). Therefore, the observed anisotropy at depth between 10 km and 30 km is unlikely due to sedimentary rocks alone unless they extend to much greater depths than expected.

Layered sill complexes have been suggested to produce strong ($> 8\%$) positive radial anisotropy at 5 – 18 km depth below silic calderas (Jiang et al., 2018), but it is usually very localized in depth and spatial extent. The strength and extent of the anisotropy observed here would require a lot of melt distributed in the Martian crust over a broad depth range. Alternatively, with Mars hosting the largest volcano of the solar system north-east of InSight, Olympus Mons, as well as Elysium Mons to the north, multiple volcanic events over broad scales could have led to an alternation of basalt layers deposit and sedimentation, which could potentially have generated radial anisotropy, though, again, it would have to extend over rather large depths. Basalt packages in a terrestrial context (for example, those associated with large igneous provinces, e.g., Planke (1994); Christie et al. (2006)) show a high degree of velocity layering, variation within which easily exceeds the 20 % perturbation required to reproduce our inferred ξ value (Baan et al., 2007). Much of this heterogeneity is associated with variations of porosity and sub-aerial fracturing (Planke, 1994) which might be expected to be reduced at greater depth in the Earth. It has been suggested, however, that on Mars pore space might persist to the observed seismic discontinuity at 20 km below the InSight lander (Knapmeyer-Endrun et al., 2021; Wieczorek et al., 2022).

Another option is the presence of layered intrusions due to a single or multiple impacts (Wilhelms & Squyres, 1984; Nimmo & Tanaka, 2005; Andrews-Hanna et al., 2008; Marinova et al., 2008). It has been proposed that the crustal dichotomy, characterized by a difference of up to 30 km in crustal thickness between the northern lowlands and the southern highlands, corresponds to a crater formed by a large impact. Such an impact could have contributed to the evolution of the proto-crust on Mars and other terrestrial planets, and could have caused internal layering within the crust. This has been shown to be the case on Earth: Latypov et al. (2019) studied the exposed impact melt

sheet of the Sudbury Igneous Complex located in Canada and showed that such ancient and compositionally primitive Hadean impact melt sheets underwent large-scale igneous differentiation.

This scenario is also consistent with the results of Li, Beghein, Lognonné, et al. (2022) who constrained crustal V_{SV} in the highlands by measuring and inverting the dispersion of minor-arc (R1) and major-arc (R2) Rayleigh waves, i.e. the wave that travels the shortest and the longest distance, respectively, along the great circle path between the epicenter and the instrument, according to ray theory. They found V_{SV} between about 3.3 km/s and 3.5 km/s in the highlands, which is greater than the crustal V_{SV} of 3.0 – 3.2 km/s found here in the lowlands, but similar to the 3.4 km/s we found for crustal V_{SH} in the lowlands (Fig. S8(b), S8(e), and S8(h)). This similarity in velocity between V_{SH} in the lowlands and V_{SV} in the highlands could indicate that the average Martian upper crustal shear wave velocities might have been originally similar in the lowlands and highlands. The impact-related horizontal melt sheets created at depth in the lowlands could have lowered the shear wave velocity in the vertical direction (i.e., V_{SV}) but left V_{SH} unchanged (Fig. S12(a)).

Finally, we cannot entirely rule out the lattice preferred orientation of anisotropic crystals in the Martian crust to explain the observed signal, as proposed for Earth’s mid-to-lower continental crust (Shapiro et al., 2004). There is currently no large underthrusting on Mars such as that seen in Tibet (Shapiro et al., 2004), and there is no evidence of large-scale crustal thinning between the lander and the event epicenter, but one could envision that anisotropic crystal fabrics could be preserved in the crust from past deformation such as an impact. If the impact melted the crust and it began to flow, fabric could have been generated over relatively large-scales and become frozen-in as rocks cooled down.

We end this discussion by pointing out that, importantly, the observed anisotropy ($\xi > 1$) in this study does not contradict the previously observed anisotropy beneath the InSight lander site (Li, Beghein, Wookey, et al., 2022). First, the anisotropy in Li, Beghein, Wookey, et al. (2022) reflects the local structure (i.e., within 30 km) beneath the lander, and this study represents the average structure along the $> 2,000$ km long path between the epicenter and the station. Second, the anisotropy in Li, Beghein, Wookey, et al. (2022) was restricted to the topmost layer (around 8 km), whereas the most re-

liably resolvable depths in the present study are 10 – 25 km. Last, but importantly, the direction of the ray path is near-vertical for the body waves in Li, Beghein, Wookey, et al. (2022) and is horizontal for the surface waves in this study. Therefore, the shear-wave velocity we constrain from the surface waves is that of a horizontally propagating shear-wave. The definitions of V_{SV} are not thus the same in both studies since the directions of the polarization of the SV waves (perpendicular to the direction of the seismic rays) are different (Fig. S12(b) – (c)). Subsequently, the $V_{SV} < V_{SH}$ anisotropy in this study does not contradict the $V_{SV} > V_{SH}$ anisotropy in Li, Beghein, Wookey, et al. (2022) and they reflect different properties of the materials in the Martian crust. In fact, the modeling done by Li, Beghein, Wookey, et al. (2022) to explain their observations used a horizontally transverse isotropic medium, which reflects the presence of azimuthal anisotropy. Here, the signal reflects radial anisotropy, also called polarization anisotropy or vertically transverse isotropy (VTI), which has hexagonal symmetry with a vertical symmetry axis. The observations beneath the lander require a medium with a higher degree of symmetry.

6 Conclusion

We analyzed the Rayleigh and Love waves generated by the largest event ever recorded on Mars. The waveforms contain surface wave energy on all three components of the seismogram at periods between 15 s and 40 s. We measured the group velocity dispersion of these waves on the vertical, radial, and transverse components of the seismogram using a multiple filter analysis. We found that Rayleigh waves group velocities increase continuously up to about 3 km/s between 15 s and 40 s period. Love wave group velocities are systematically larger, between about 3.2 km/s and 3.5 km/s in the 15 – 40 s period range. Because of a discrepancy between the Rayleigh wave group velocities in the transverse and vertical directions at periods greater than 30 s, we only used data up to 30 s.

Isotropic inversions of the Love wave data were unable to explain the Rayleigh wave data and vice versa. The Love and Rayleigh wave dispersion curves were thus inverted jointly for shear wave velocities and radial anisotropy. We found a well-resolved signal in the crust, with $V_{SH} > V_{SV}$ ($\xi > 1$) between 10 km and 25 km depth.

LPO of anisotropic crystals inside the crust is difficult to reconcile with our current knowledge of the martian crust as it would require a large-scale shear mechanism

360 such as crustal thinning or underthrusting. However, we could be seeing frozen-in fab-
361 ric due to melting followed by flow (and thus LPO) in the crust after an impact, for in-
362 stance. The observed anisotropy may alternatively be apparent rather than intrinsic. We
363 rule out cracks as a possible explanation for this signal since vertical cracks would yield
364 $V_{SH} < V_{SV}$. Horizontal layering of sediments is unlikely to explain the signal as well
365 because they would not extend to great enough depths. Our preferred explanations in-
366 clude an alternation of isotropic basalt layers deposit and sedimentation caused by vol-
367 canic eruptions, or the presence of isotropic layered intrusions due to an impact.

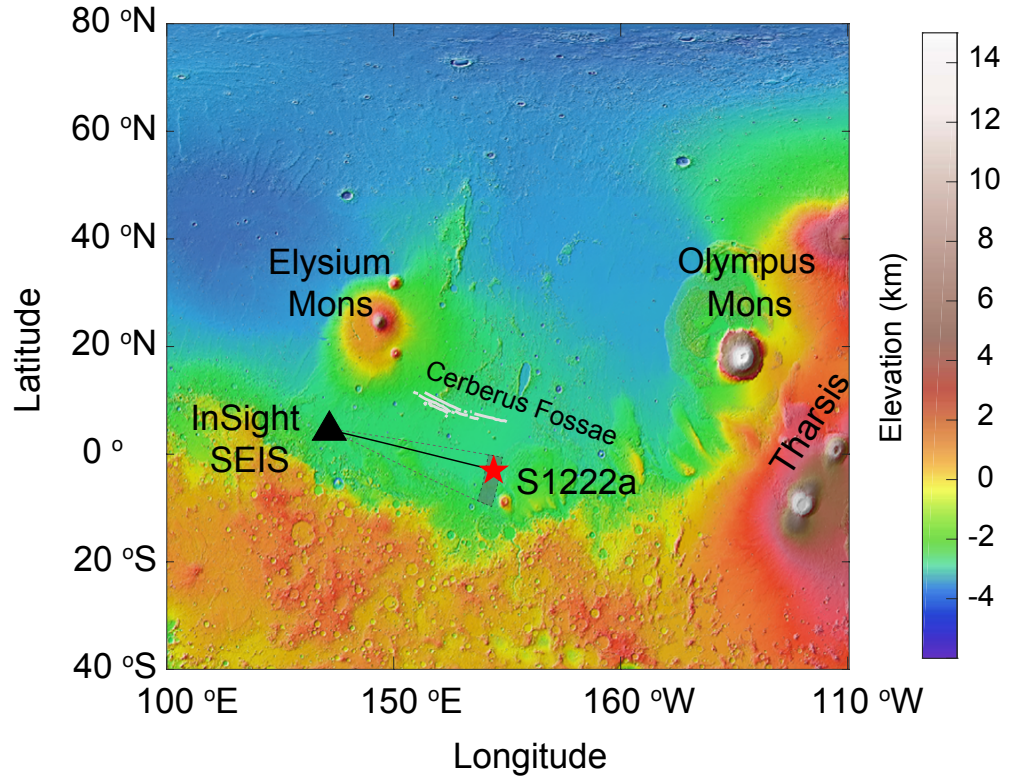


Figure 1. Location of event S1222a (red star) and SEIS (black triangle). The grey shaded area represents the uncertainties in back azimuth and epicentral distance and the three black curves are possible great circle paths of the surface waves.

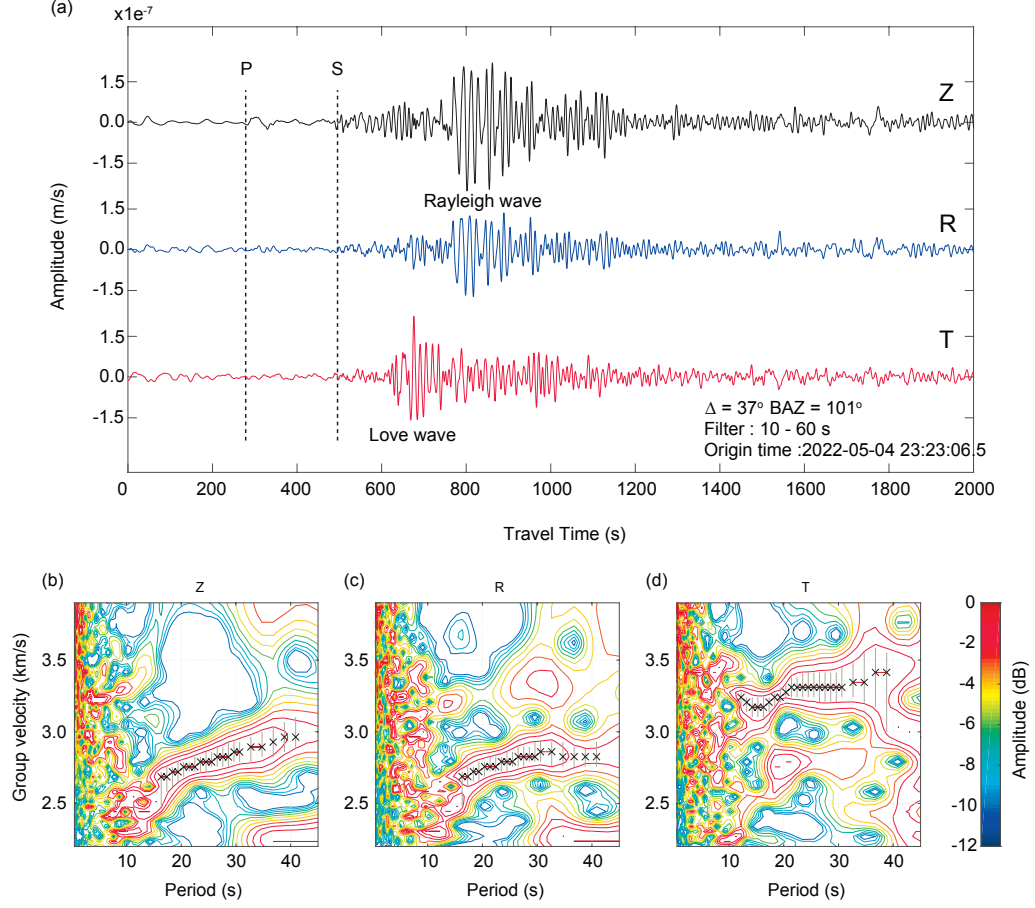


Figure 2. (a) Deglitched waveforms on the to Z, R, T coordinates. A second order, zero-phase shift bandpass filter was applied between 10 s and 60 s period. (b)-(d) Multiple filter analysis of the vertical (a), radial (b), and tangential (c) components of the VBB.

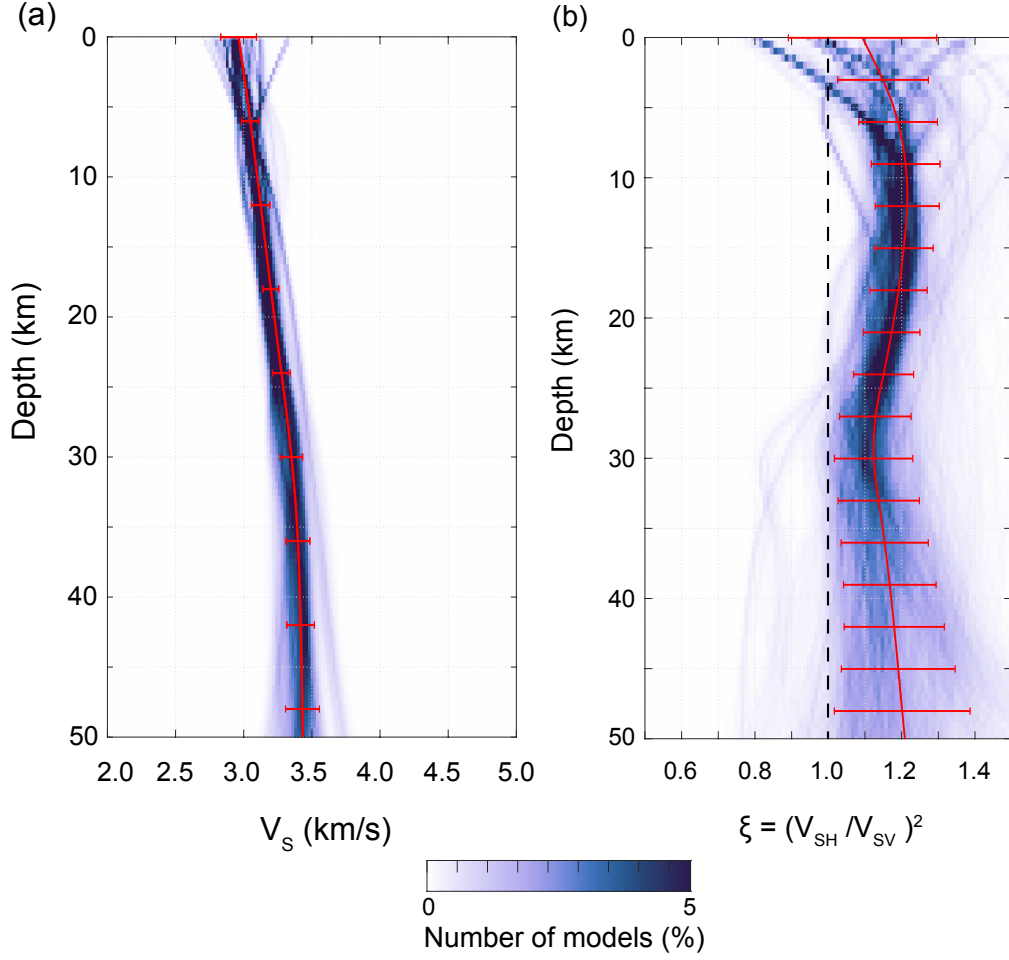


Figure 3. V_S (a) and ξ (a) models resulting from Method 1 applied to the measured dispersion curves. The vertical black dashed line in (b) represents an isotropic model for reference. The red horizontal bars are the uncertainties calculated by averaging the variance obtained by each of the three methods.

7 Open Research

Data and materials availability: The InSight event catalogue (Clinton et al., 2021; Ceylan et al., 2022; InSight Marsquake Service, 2022a, 2022b) and waveform data are available from the IRIS-DMC, NASA-PDS (InSight Mars SEIS Data Service, 2019a) and IGP data center (InSight Mars SEIS Data Service, 2019b).

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The authors declare that they have no competing interests.

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