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Modelling seismic recordings of high-frequency guided infrasound on Mars

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Key Points:

- We analytically compute the dispersion relationship of guided infrasound in 1D atmospheric models from Mars and Earth;
- We model the ground response to guided infrasound to fit the corresponding seismic observation due to meteorite impacts on Mars;
- We utilize the seismic recording due to guided infrasound to constrain the sub-surface structures.

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Abstract

NASA’s InSight mission records several high-frequency (>0.5 Hz) dispersive seismic signals on Mars. These signals are due to the acoustic-to-seismic coupling of infrasound generated by the entry and impact of meteorites. This dispersion property is due to infrasound propagating in a structured atmosphere, and we refer to this dispersive infrasound as guided infrasound. We propose to model the propagation of guided infrasound and the seismic coupling to the ground analytically; we use a 1D layered atmosphere on a three-layer solid subsurface medium. The synthetic ground movements fit the observed dispersive seismic signals well and the fitting indicates the regolith beneath InSight is about 40-m in thickness. We also examine and validate the previously-published subsurface models derived from InSight ambient seismic vibration data.

Plain Language Summary

Under particular weather conditions, the Martian atmosphere displays a special sound-wave velocity profile, where the wave velocity becomes larger with increasing altitude within a few hundred meters. When an infrasound signal - a low-frequency (<20 Hz) sound wave inaudible to humans - propagates through such a structure, the infrasound exhibits dispersion : its propagation velocity depends on its frequency. We refer to such infrasound as guided infrasound. Guided infrasound can deform the ground, and have been recorded by the seismometer of NASA’s InSight mission on the Martian surface. We propose to model these recordings using the physics of sound waves traveling above a compliant solid ground. We show that our modelling results can fit well the seismic recordings of guided infrasound on Mars. We apply our modelling to the subsurface models from a different InSight seismic observation to check if these models can explain our seismic recordings. This modelling constitutes a new tool to investigate the subsurface structure of Mars, and is also useful for the investigation of Titan and Venus.

1 Introduction

NASA’s InSight (Interior Exploration using Seismic Investigations, Geodesy and Heat Transport) mission landed on the Martian surface in November 2018 and has since been conducting geophysical and meteorological observation (Banerdt et al., 2020). To achieve its objectives, InSight is equipped with a Very Broad Band (VBB) and a Short Period (SP) seismometer, which together constitute the SEIS (Seismic Experiment for Internal Structure) instrument (Lognonné et al., 2019). SEIS is operated in combination with a weather station, Auxiliary Payload Sensor Suite (APSS) including an atmospheric pressure sensor and wind and temperature sensors, to perform meteorological observation (Banfield et al., 2019). Due to power issues appearing in the second Martian year of the mission, SP and APSS have become temporarily unavailable, and VBB has been kept on most of the time. Thus only the VBB seismic data is available for analyzing the seismic events in this study.

The ground motion recorded by InSight originates from different types of sources, most of which are marsquakes (e.g. Giardini et al., 2020) or atmospheric seismic events like pressure drops (e.g. Lognonné et al., 2020). The recent seismic recordings provides a new type of seismic events, a dispersive wave train following a typical very-high-frequency (VF) marsquake (Clinton et al., 2021), where a dispersive wave train means that the wave velocity, also arrival time, depends on frequency. This wave train appears about a few hundred of seconds after the P arrivals, such as in events S0793a, S0981c, and S0986c. Based on analysis of seismic arrival times and satellite imagery, Garcia et al. (2022) demonstrate that these events are generated by meteorite impacts on the Martian surface (Figure 1). Thus, the high-frequency seismic energy arriving before the dispersive waves is due to the meteorite cratering process (Figure 2a).

73 The meteors not only generate the craters but also interact with the Martian at-
74 mosphere during entry and impacting, which generates infrasound, i.e. acoustic waves
75 with a frequency lower than 20 Hz (Figure 2). The infrasound propagation medium - the
76 atmosphere - can exhibit a particular structure, where the infrasound propagation ve-
77 locity is smaller near the ground surface than at higher altitudes (about a few hundred
78 meters). In such structures, multiple infrasound propagation paths interfere with each
79 other, and the interference generates dispersion (Herrin et al., 2006; Negraru & Herrin,
80 2009), similar to the mechanism of Love waves in seismology (e.g. Aki & Richards, 2002).
81 This infrasound velocity model in Earth is referred to as nocturnal boundary layer in acoustic-
82 wave literature (e.g. Waxler, 2004), since such a model is usually generated when the
83 temperature on the ground surface decreases at night, leading to a cooling of the lower
84 atmosphere. This phenomenon is common on Mars due to the quick cooling down of the
85 Martian surface at night and/or high-altitude winds (e.g. Garcia et al., 2017). In this
86 study, we refer to this atmospheric structure as a waveguide, and we refer to the inter-
87 ferred infrasound waves as guided infrasound. Guided infrasound can be simulated nu-
88 merically by solving the acoustic wave equations (e.g. Garcia et al., 2017; Martire et al.,
89 2020). This numerical simulation approach can address complicated atmospheric mod-
90 els like a laterally heterogeneous atmosphere with winds, but is computationally expen-
91 sive. One can also model the guided infrasound waveform analytically by calculating the
92 phase/group velocity (i.e. dispersion) within a laterally-homogeneous two-layer atmo-
93 spheric model (Negraru & Herrin, 2009). This analytical approach is much faster and
94 less computationally expensive than the numerical simulation, and is therefore well suited
95 to explore different atmospheric parameters and their associated guided infrasound. How-
96 ever, former studies remain limited to two-layer atmospheric models. In this study, we
97 extend this analytical approach to a multiple-layer model using the propagation matrix
98 method described in seismic surface-wave studies (e.g. Aki & Richards, 2002). Note that
99 in this study we focus on the high-frequency guided infrasound, while on Earth, the low-
100 frequency (<0.02 Hz) guided infrasound also exists (e.g. Pekeris, 1948; Press & Harkrider,
101 1962; Harkrider, 1964).

102 We use the VBB seismic data (InSight Mars SEIS Data Service, 2019a, 2019b) to
103 study guided infrasound observed on Mars, because infrasound propagates with atmo-
104 spheric perturbations and the perturbations deform the ground (e.g. Sorrells, 1971; Tan-
105 imoto & Wang, 2019). One has observed this type of coupling due to atmospheric pres-
106 sure drops through seismic recordings on both Earth and Mars (e.g. Lorenz et al., 2015;
107 Lognonné et al., 2020; Kenda et al., 2020). Furthermore one also observes infrasound de-
108 forming Earth’s ground surface due to acoustic sources such as volcanic activities (e.g.
109 Ichihara et al., 2012), meteors (e.g. Edwards et al., 2008), and ground surface explosions
110 (e.g. Gibbons et al., 2007; Schneider et al., 2018). Conversion of an atmospheric pres-
111 sure perturbation into ground deformation is called compliance, and the intensity of com-
112 pliance is determined by the subsurface structure and the propagation velocity of the per-
113 turbation (e.g. Ewing et al., 1957; Sorrells et al., 1971; Ben-Menahem & Singh, 2012).
114 We detail computation of compliance in Section 3.

115 We propose to analytically model the seismic recordings due to guided infrasound,
116 and we refer to the recordings as chirps. We demonstrate computation of the guided in-
117 frasound phase and group velocities in multiple-layer atmospheric models (Section 2).
118 We then introduce the compliance and our subsurface velocity model (Section 3). We
119 combine the guided infrasound and the compliance to generate a synthetic chirp; we use
120 the synthetic chirp to fit the observed ones (Section 4). We finally discuss the implica-
121 tion of our modelling to the previously-published subsurface models and the infrasound
122 propagation in the Martian atmosphere (Section 5). Our research can aid the investi-
123 gation of atmospheric and subsurface properties, not only on Mars and Earth but also
124 on other bodies with atmosphere such as Titan and Venus.

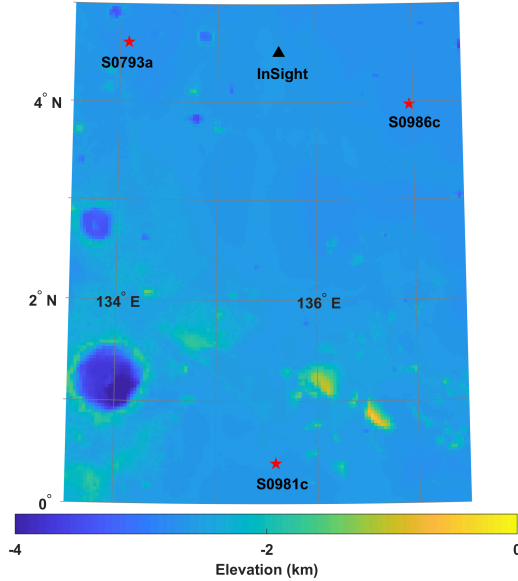


Figure 1. Topography map (Smith et al., 2001) of the Martian surface around InSight and three impact locations (S0793a, S0981c, and S0986c). The impact locations are constrained by CTX images (Garcia et al., 2022).

2 Modelling Guided infrasound

2.1 Theory

We solve the dispersion of the guided infrasound in a multiple-layer atmospheric model theoretically. We present an example of the Martian guided infrasound in this section and an Earth example in Appendix A. We assume the atmosphere to be adiabatic, which leads to the following 2D governing equations:

$$-\partial_z p = \rho D_t v_z, \quad (1)$$

$$-\partial_x p = \rho D_t v_x, \quad (2)$$

$$-D_t p = K(\partial_z v_z + \partial_x v_x), \quad (3)$$

where ∂_z denotes the partial derivative with respect to altitude (z), p is the atmospheric pressure perturbation conveyed by the infrasound wave, ρ is the air density, and K is the incompressibility; v_z and v_x are the particle velocities in the vertical and horizontal directions, respectively. D_t denotes the material derivative with respect to time. Indeed, the advection of momentum cannot be ignored here, as the horizontal wind velocity (w_x) can reach up to about 15 m/s on the Martian surface. The effect of w_x is made explicit by rewriting the above equations as:

$$-\partial_z p = \rho(\partial_t + w_x \partial_x) v_z, \quad (4)$$

$$-\partial_x p = \rho(\partial_t + w_x \partial_x) v_x, \quad (5)$$

$$-(\partial_t + w_x \partial_x) p = K(\partial_z v_z + \partial_x v_x). \quad (6)$$

Since in the three impact events, the cross winds deviate the infrasound propagation direction by less than 5 deg (Garcia et al., 2022), we neglect the cross wind effect and project the total wind speed to the guided infrasound propagation direction to achieve w_x (Table S1, S2, and S3). Besides the above equations, a set of boundary conditions is also needed. We adopt a rigid ground surface, such as:

$$v_z(z = 0) = 0. \quad (7)$$

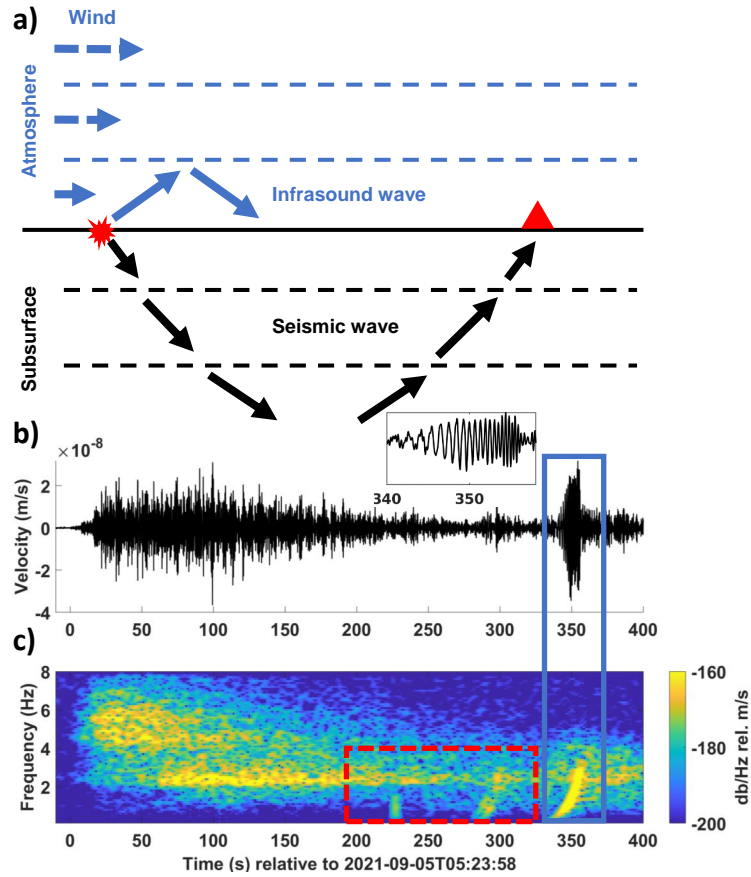


Figure 2. Illustration of the S0986c event (a) and data (b, c). (a) The meteor of the S0986c impact interacts with the Martian subsurface and atmosphere, generating seismic waves (from 0 to about 200 s in (b)) and infrasound (after 200 s in (b)), respectively. The vertical scales of the atmosphere and the subsurface are in hundred meter and kilometer, respectively. (b) The S0986c VBB vertical-component data. The data is bandpass filtered between 0.1 and 8 Hz. The insert shows a zoom of the chirp in the blue box. (c) Spectrogram of the data in (b). The blue box indicates the chirp spectrogram. The red box indicates other arrivals of infrasound energy. The time axis is in Coordinated Universal Time (UTC).

153 Note that one may choose an elastic ground-surface boundary condition and then the
 154 right hand side of the above equation would not be zero anymore (e.g. Waxler, 2002).
 155 We justify our rigid boundary condition in Section 5.3. We use the radiation boundary
 156 condition for the top boundary. The 2D geometry assumed here is a good approxima-
 157 tion for a real 3D geometry only if the cross-wind can be neglected (e.g. Nijs & Wape-
 158 naar, 1992). Other second-order factors are not considered here, such as the effect of static
 159 pressure or gravity (e.g. Pierce, 1990).

160 We compute the guided infrasound phase velocities by solving the above equations
 161 (Equation 4, 5 and 6) and the boundary conditions (e.g. Equation 7). One can adopt
 162 normal mode expansion to solve the equations, and Assink (2012) present a complete
 163 review of this approach mathematically. However, this approach is not designed specif-
 164 ically for guided infrasound and is built on an elastic boundary condition, instead of our
 165 rigid one (Equation 7). Thus, we propose a simpler method. We assume that guided in-
 166 frasound propagates horizontally as a plane wave:

$$p = P(\omega, z) \exp(i(\omega t - kx)), \quad (8)$$

$$v_z = V_z(\omega, z) \exp(i(\omega t - kx)), \quad (9)$$

$$v_x = V_x(\omega, z) \exp(i(\omega t - kx)), \quad (10)$$

167 where ω is the angular frequency, z is altitude, t is the propagation time, k is the hor-
 168 izontal wavenumber, and x is the propagation distance. Note that ω/k gives the phase
 169 velocity. Based on Equation 5, 8, and 10, we notice that

$$V_x = \frac{kP}{\rho(\omega - w_x k)}. \quad (11)$$

170 We then rewrite the governing equations in a matrix form as

$$\partial_z \begin{bmatrix} P \\ V_z \end{bmatrix} = \begin{bmatrix} 0 & -i(\omega - w_x k)\rho \\ -i\omega/\alpha^2/\rho + iw_x k/\rho/\alpha^2 + ik^2/(\omega - w_x k)/\rho & 0 \end{bmatrix} \begin{bmatrix} P \\ V_z \end{bmatrix}, \quad (12)$$

171 where α is the infrasound velocity, where $\alpha^2 = K/\rho$. Note that Press and Harkrider
 172 (1962) and Nijs and Wapenaar (1990) have achieved similar equations as Equation 12.
 173 Press and Harkrider (1962) study the low-frequency (<0.02 Hz) guided infrasound con-
 174 sidering gravity, and Nijs and Wapenaar (1990) study acoustic wave propagation and do
 175 not consider the guided infrasound.

176 We use the propagation matrix method (e.g. Aki & Richards, 2002) to solve Equa-
 177 tion 12. This equation is in the form of $\partial_z \vec{f} = \mathbf{A} \vec{f}$, where \vec{f} is normally referred to as
 178 eigenfunctions and \vec{f} here contains P and V_z . The propagation matrix is defined as $\mathbf{M}(z, z') =$
 179 $\exp[(z - z')\mathbf{A}]$ between two depths, z and z' . We use \mathbf{M} to calculate the eigenfunctions
 180 at z from z' as

$$\vec{f}(z) = \mathbf{M}(z, z') \vec{f}(z'). \quad (13)$$

181 We observe that Equation 13 intuitively satisfies Equation 12. For a 1D atmospheric model,
 182 with a right phase velocity (i.e. ω/k) at frequency ω , we can calculate the eigenfunctions
 183 at all altitudes, and the eigenfunction values on the ground surface would satisfy the bound-
 184 ary condition (Equation 7). The computation is detailed in Section S1. Based on the phase
 185 velocity, we can also compute the group velocity as the variation of ω over the variation
 186 of k (e.g. Aki & Richards, 2002).

187 2.2 S0986c guided infrasound

188 We use an atmospheric model of Mars to illustrate the eigenfunctions and the group
 189 velocities of the S0986c guided infrasound (Figure 3). We adopt the atmospheric model
 190 parameters (acoustic-wave velocity, wind speed, air density, and altitude) from the Mars

191 Climate Database (MCD, Millour et al., 2018). We project the wind speed along the back-
 192 azimuth of SEIS with regards to the S0986c impact location (Figure 1 and Table S3);
 193 the cross wind of this event atmospheric model does not obviously affect the infrasound
 194 propagating from the impact location to InSight (Garcia et al., 2022). Note that MCD
 195 models the Martian climate at a global scale and could be biased at local scale like our
 196 cases. Thus we modify the acoustic-wave velocity by fitting the synthetic group veloc-
 197 ity to the measurement from the chirp (Garcia et al., 2022). We add (subtract) a con-
 198 stant value to (from) acoustic-wave velocities at all altitudes, which moves the whole syn-
 199 thetic group-velocity curve up (down) but does not change the synthetic group-velocity
 200 shape (Figure 3c). The sum of the projected wind speed and the modified acoustic-wave
 201 velocity is called the effective velocity (Figure 3a). The effective velocity of this event
 202 increases gradually with altitude until about 500 m and then decreases slightly. We dis-
 203 cretize the wind speed and the acoustic-wave velocity to form a 1D layered model, and
 204 then compute the phase velocity (Figure 3c) and the atmospheric pressure eigenfunc-
 205 tions (P , Figure 3b). The pressure eigenfunctions are real valued and maximal on the
 206 ground surface. We also observe that a high-frequency (3 Hz) eigenfunction decays faster
 207 than a low frequency one (1 Hz) with increasing altitude. The pressure eigenfunction rep-
 208 represents the amplitudes (P in Equation 8) of the guided infrasound at different altitudes.
 209 For example, in this case, we would receive a lower-amplitude guided infrasound with
 210 an atmospheric-pressure sensor at 500 m altitude (like a balloon) compared to a sensor
 211 on the ground surface. The eigenfunctions also represent the excitation amplitude of the
 212 guided infrasound due to the source-receiver reciprocity (e.g. Landau & Lifshitz, 2013;
 213 Aki & Richards, 2002). Therefore an infrasound source at 500 m would generate weaker
 214 infrasound compared to a source on the ground surface. We use the phase velocity later
 215 in our computation of compliance (Section 3) and our chirp modelling (Section 4).

216 2.3 Higher-mode guided infrasound

217 We present the fundamental-mode guided infrasound in the above. The fundamental-
 218 mode represents the lowest-phase-velocity root in solving Equation 12 at each frequency,
 219 while the higher-velocity roots may also exist (Figure S1) and are referred to as the higher
 220 modes. The fundamental mode usually dominates guided infrasound in observations (e.g.
 221 Negraru & Herrin, 2009). Thus in the group-velocity measurement of S0981c and S0986c,
 222 we only observe the trends corresponding to the fundamental modes (Figure S1b and
 223 S1c).

224 We also demonstrate the domination of the fundamental-mode by computing the
 225 contribution of both the fundamental and higher modes to the pressure on the ground
 226 surface as $P^2(z=0)/\int P^2(z)dz$. The pressure eigenfunction (P) is real valued if the cor-
 227 responding mode is a trapped mode (e.g. Lognonné et al., 1998; Chakravarthy, 2008).
 228 However, at some frequencies, the eigenfunctions become complex valued, and the imag-
 229 inary parts represent the energy leakage of the guided infrasound to the top halfspace
 230 (e.g. Press & Harkrider, 1962; Radovich & De Bremaecker, 1974). Thus we only use the
 231 real part of the eigenfunctions on the ground surface in computing the contributions (Fig-
 232 ure S1). We observe that for S0981c and S0986c, the contribution of the fundamental
 233 modes are at least one magnitude (a factor of ten) larger than the higher modes at each
 234 frequency. However, for S0793a, the 1st-higher mode contributes more than the funda-
 235 mental mode. We investigate this phenomenon in detail in Section 5.2.

236 3 Compliance: acoustic-to-seismic conversion

237 Atmospheric pressure perturbations, e.g. caused by wind or infrasound, can deform
 238 the ground at shallow depths on planets with atmosphere (e.g. Sorrells, 1971; Ben-Menahem
 239 & Singh, 2012), like the atmospheric noise recorded by SEIS on Mars (Lognonné et al.,
 240 2020; Garcia et al., 2020; Stutzmann et al., 2021). Furthermore, when the infrasound hor-

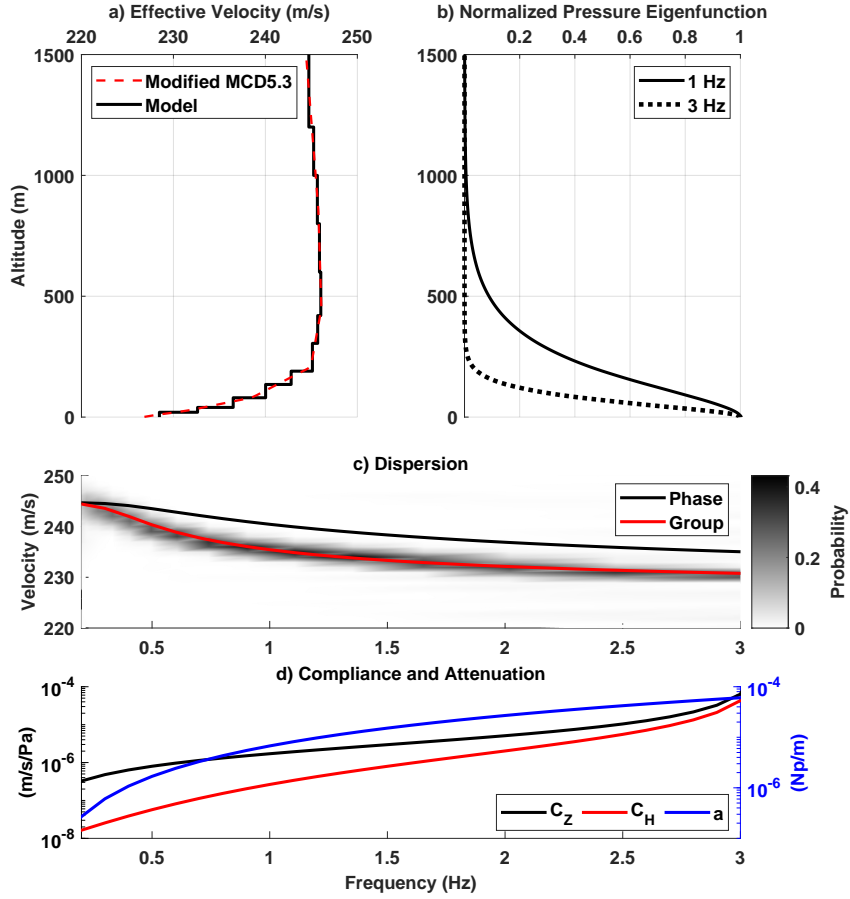


Figure 3. The infrasound velocity model (a), eigenfunctions (b), dispersion (c) and compliance (d) for S0986c. (a) We compute the infrasound velocity and the wind speeds from MCD for S0986c (Millour et al., 2018) and subtract 4 m/s from the velocity at all altitudes (red dashed line). The black line represents our 1D layered model. (b) Two different-frequency pressure eigenfunctions are normalized by each maximum value for visualization. (c) The gray background is the group-velocity measurement of the S0986c chirp (Garcia et al., 2022). (d) The vertical (black) and horizontal (blue) compliance is based on the phase velocity (c) and a subsurface velocity model (Table 1). The infrasound attenuation (blue) is from Bass and Chambers (2001) for 200 K temperature.

241 izontal apparent velocity is equal to the seismic-wave velocities of the subsurface, the in-
 242 frasound would convert into seismic waves (e.g. Ewing et al., 1957; Langston, 2004). To
 243 summarize these different types of acoustic-to-seismic coupling, Edwards et al. (2008)
 244 list the possible conversion scenarios for homogeneous and isotropic elastic subsurface
 245 media. Note that in this study, we mainly focus on the so-called normal coupling, i.e.
 246 atmospheric pressure perturbations deforming the ground surface without generating seis-
 247 mic waves, but the following theory remains applicable to all the scenarios mentioned
 248 above. Due to the frequency band we use (here from 0.5 to about 2 Hz), we only con-
 249 sider the compliance in this study and ignore other effects like tilt (Garcia et al., 2020).

250 Compliance is the amplitude relationship between the pressure perturbation and
 251 the ground deformation. Compliance is determined by the subsurface structure and the
 252 propagation velocity of the pressure perturbation in the atmosphere overlaying the sub-
 253 surface (e.g. Sorrells, 1971; Ewing et al., 1957; Ben-Menahem & Singh, 2012), based on
 254 the assumption that the perturbation propagates like a plane wave. Note that the plane-
 255 wave assumption even holds for a complex pressure wavefield, since the complex wave-
 256 field can be decomposed into plane waves (e.g. Kenda et al., 2017). In the subsurface
 257 media, the ground motion and normal stress (τ_{zz}) are defined as:

$$258 \quad u_z(z) = U_z(\omega, z) \exp(i(\omega t - kx)), \quad (14)$$

$$259 \quad u_x(z) = iU_x(\omega, z) \exp(i(\omega t - kx)), \quad (15)$$

$$260 \quad \tau_{zz}(z) = T_{zz}(\omega, z) \exp(i(\omega t - kx)), \quad (16)$$

262 where u_z and u_x are the vertical and horizontal ground velocities, respectively; ω , k , t ,
 263 and x are the same as in guided infrasound (Equation 8) while z here is the depth be-
 264 low the surface. The i in front of U_x in Equation 15 represents the $\pi/2$ phase shift be-
 265 tween the horizontal and vertical components of the ground motion (e.g. Sorrells, 1971).
 266 We then write compliance as the amplitude ratio of the vertical (horizontal) ground ve-
 267 locities over the atmospheric pressure perturbation:

$$268 \quad C_z = \frac{u_z(z=0)}{p(z=0)} = \frac{U_z(\omega, z=0)}{P(\omega, z=0)} = -\frac{U_z(\omega, z=0)}{T_{zz}(\omega, z=0)}, \quad (17)$$

$$269 \quad C_x = \frac{u_x(z=0)}{p(z=0)} = \frac{iU_x(\omega, z=0)}{P(\omega, z=0)} = -\frac{iU_x(\omega, z=0)}{T_{zz}(\omega, z=0)}, \quad (18)$$

271 where $T_{zz}(z=0) = -P(z=0)$ (Equations 8 and 16) implies the continuity of nor-
 272 mal stress on the ground surface. The minus sign is due to the different sign conventions
 273 between atmospheric studies and seismology. In acoustics wave studies (Section 2), the
 274 atmospheric pressure compressing the ground surface is defined as being positive, i.e. ex-
 275 erting a force in the vertically downward direction on the ground surface. However in
 276 seismology, a positive normal stress acting on a surface corresponds to a traction in the
 277 outward normal direction, i.e. the vertically upward direction on the flat ground surface.
 278 Note that here we refer to the vertical ground velocity as u_z , not to be mistaken with
 279 the vertical particle velocity in the atmosphere, i.e. v_z in Section 2.

280 To calculate the compliance values, we use the equation of motion and Hooke's law
 281 in 1D media (e.g. Aki & Richards, 2002, Chapter 7.2) and the zero-shear-stress bound-
 282 ary condition on the ground surface:

$$283 \quad \tau_{zx} = 0. \quad (19)$$

284 Our computation is similar to Tanimoto and Wang (2019) where one does not assume
 285 the propagation velocity of pressure perturbation much slower than the shear-wave ve-
 286 locity of the subsurface medium. In this study, we use a three-layer velocity model sim-
 287 plified from the shallow (<100 m) geological structure under InSight (e.g. Warner et al.,
 288 2022). The first layer is made of thin soft regolith, as suggested by the analyses of the
 289 Martian atmospheric pressure drops (Kenda et al., 2020; Onodera, 2022) and of the ham-
 290 mering of InSight's Heat Flow and Physical Properties (HP³) instrument (Lognonné

Table 1. Parameters for a three-layer subsurface. The first two layers are regolith. In the first two layers, we calculate V_P from V_S based on the 0.22 Poisson’s ratio (e.g. Morgan et al., 2018) ; we compute the density applying Gardner’s empirical relationship (Gardner et al., 1974) to V_P .

Layer number	$V_P(m/s)$	$V_S(m/s)$	Density (kg/m^3)	Thickness (m)
1	117	70	1019	0.6
2	384	230	1372	40
3	3000	1700	2760	∞

291 et al., 2020). This surface layer is interpreted as fine-sand-dominated regolith. The sec-
 292 ond layer possesses a higher wave velocity than the first layer and represents coarse re-
 293 golith (e.g. Warner et al., 2017). The third layer (i.e. halfspace) corresponds to the bedrock,
 294 composed of fractured basalt (e.g. Morgan et al., 2018). We use this three-layer model
 295 to compute compliance in the waveform modelling (Section 4). We achieve the param-
 296 eters of the second layer through a waveform fitting in Section 4. Note that this model
 297 is simple and may not reflect the complexity of the real subsurface under InSight. We
 298 discuss possible improvement to this model in Section 5.4.

299 We notice that the difference of vertical velocities on the ground surface between
 300 the atmosphere and the subsurface give rise to a contradiction. In Section 2, we assume
 301 the vertical atmosphere velocity on the ground surface to be zero (Equation 7). The par-
 302 ticle velocity should be continuous at the fluid-solid (i.e. atmosphere-ground) interface,
 303 and thus $u_z(z = 0) = 0$. However, a nontrivial compliance requires the vertical move-
 304 ments of the ground surface to be non-zero, i.e. $u_z(z = 0) \neq 0$. We address this con-
 305 tradiction in Section 5.3. Note that such contradiction does not exist for the horizon-
 306 tal velocities, since the horizontal velocity of the ground surface is not necessarily con-
 307 tinuous with the horizontal atmospheric particle velocity.

308 4 Waveform Forward modelling

309 We model synthetic chirp, the seismic recording due to guided infrasound, by combin-
 310 ing the theories of guided infrasound (Section 2) and compliance (Section 3). The far-
 311 field synthetic ground velocity recordings of chirps (u) in the time domain are written
 312 as:

$$u_z = \mathcal{F}^{-1} \left[S(\omega) \exp(-ikx - ax) \frac{\exp(-i\pi/4)}{\sqrt{kx\pi/2}} \frac{P^2(\omega, 0)}{\int P^2(\omega, z) dz} C_z(\omega) \right], \quad (20)$$

$$u_x = \mathcal{F}^{-1} \left[S(\omega) \exp(-ikx - ax) \frac{\exp(-i\pi/4)}{\sqrt{kx\pi/2}} \frac{P^2(\omega, 0)}{\int P^2(\omega, z) dz} C_x(\omega) \right], \quad (21)$$

313 where \mathcal{F}^{-1} denotes the inverse Fourier transform, and S is the source time function of
 314 guided infrasound. k is guided infrasound horizontal wavenumber and is from our cal-
 315 culation of the guided infrasound phase velocity (Section 2). x is infrasound travelling
 316 distance from the infrasound source to InSight (Table 2). a is the intrinsic attenuation
 317 coefficient for infrasound (Figure 2d and Bass & Chambers, 2001). $\exp(-i\pi/4)/\sqrt{kx\pi/2}$
 318 is from the far-field approximation of a 3D cylindrical wave (e.g. Landau & Lifshitz, 2013).
 319 $P^2(\omega, 0)/\int P^2(\omega, z) dz$ is the normalized pressure eigenfunction, representing the source
 320 excitation and sensor receiving on the ground surface. C_z (C_x) is the vertical (hori-
 321 zontal) compliance (Section 3). We use the three-layer subsurface velocity model (Table 1)
 322 in computing the compliance, where the model is from simplification of the shallow ge-
 323 ological structure under InSight (e.g. Warner et al., 2022). We benchmark our modelling

Table 2. The origin time estimates and the geometry information of the three impact events relative to InSight (Garcia et al., 2022).

Event	Estimated origin time (UTC)	Distance (km)	Backazimuth (deg)
S0793a	2021-2-18T19:36:06	91.1	274.2
S0981c	2021-8-31T04:3:13	243.6	179.2
S0986c	2021-9-5T05:23:44	85.1	111.6

324 approach against a numerical simulation software, SPECFEM2D-DG (Martire et al., 2020),
 325 in Appendix B.

326 We use the synthetic chirp to fit the one due to an impact. We assume that the
 327 infrasound source generated by the meteor impacting is impulsive and thus the source
 328 time function (S in Equation 20 and 21) is a delta function in the time domain and a
 329 constant in the frequency domain. The source location is at the same location as the im-
 330 pact crater. Note that the source time function generated by a meteorite interacting with
 331 the Martian atmosphere in the impact is worth further investigation, but is beyond the
 332 scope of this study. Since the infrasound source is on the ground surface, we mainly ob-
 333 serve the guided infrasound on InSight (Garcia et al., 2022).

334 We use the observed chirps from two events (S0981c and S0986c) in the fitting. For
 335 each event, we use the origin time estimate from the seismic arrival time (Table 2, Gar-
 336 cia et al., 2022). The infrasound source is at the crater associated to the events. The satel-
 337 lite images of these craters provide the exact distances and backazimuth (Table 2). Note
 338 that these distances and backazimuth are matched closely by analysis of the two seis-
 339 mic recordings, e.g. body-wave arrival times and chirp polarization (Garcia et al., 2022).
 340 We focus on the vertical and radial components, where the radial direction is parallel
 341 to a great path from the seismic source to InSight. For the seismic recordings of the two
 342 events, we rotate the north and east components to the radial direction based on the back-
 343 azimuth (Table 2). We apply a bandpass filter to the components of each chirp. For each
 344 chirp, we choose the filter band to have high signal-to-noise ratios and to avoid spectral
 345 anomaly like the sharp amplitude drop at 2.3 Hz in the S0986c vertical component (Fig-
 346 ure 4). In the following waveform fitting, we refer to the filtered data as real data. A chirp
 347 also exists in the S0793a recording. However, that chirp displays complex properties (e.g.
 348 higher-mode), which deserve to be discussed independently (Section 5.2).

349 We compute the misfit between the synthetic and real chirps like

$$\chi = \sum_j \frac{1}{T} \int_T \left| \frac{u_z(s_j)}{\max(|u_z(s_j)|)} - \frac{u_z^o(s_j)}{\max(|u_z^o(s_j)|)} \right| + \left| \frac{u_x(s_j)}{\max(|u_x(s_j)|)} - \frac{u_x^o(s_j)}{\max(|u_x^o(s_j)|)} \right| dt \quad (22)$$

350 where s_j represents each event; the integral is done in the time window of each chirp (Fig-
 351 ure 4) and T is the time window length for each chirp. We fix the first and third layers
 352 of the model, and only vary the V_S and the layer thickness of the second layer to find
 353 the misfit minimum (Figure 5). The S0981c misfit mainly varies with the layer thickness
 354 but does not change obviously with V_S when the thickness is less than 40 m (Figure 5a).
 355 The S0986c misfit presents a sloping area (Figure 5b) where the different combinations
 356 of the two parameters give similar misfits, which is known as trade-off in the geophys-
 357 ical inverse theory. In the total misfit of the two events (Figure 5c), the V_S and the layer
 358 thickness from the minimum provide good waveform fitting between the synthetic and
 359 real chirps (Figure 4). We discuss this model in the geological context in Section 5.1.

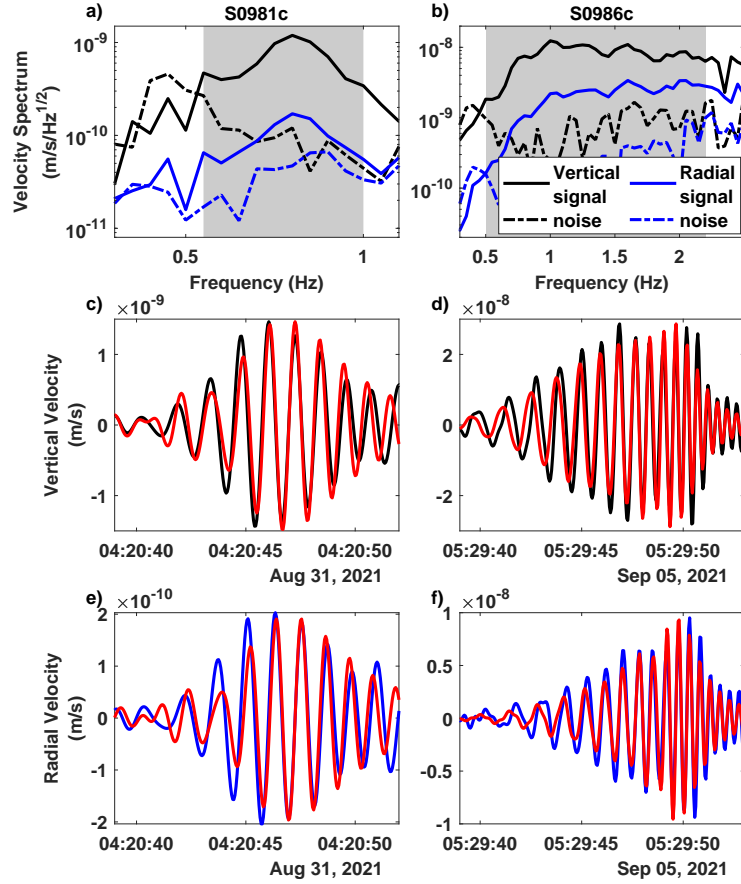


Figure 4. The vertical- and radial-component amplitude spectra of the chirps and noise (a,b) and the waveform fitting of the chirps in the time domain (c,d,e,f). The black and blue waveforms are the observed data and the red waveforms are the synthetic data. For each event, the signal spectra are from the time window used in the second and third rows, and the noise spectra are from the same-window-length recordings before the chirp. The gray areas indicate the frequency bands used in filtering the chirps, from 0.55 to 0.9 Hz for S0981c and from 0.5 to 2.2 Hz for S0986c.

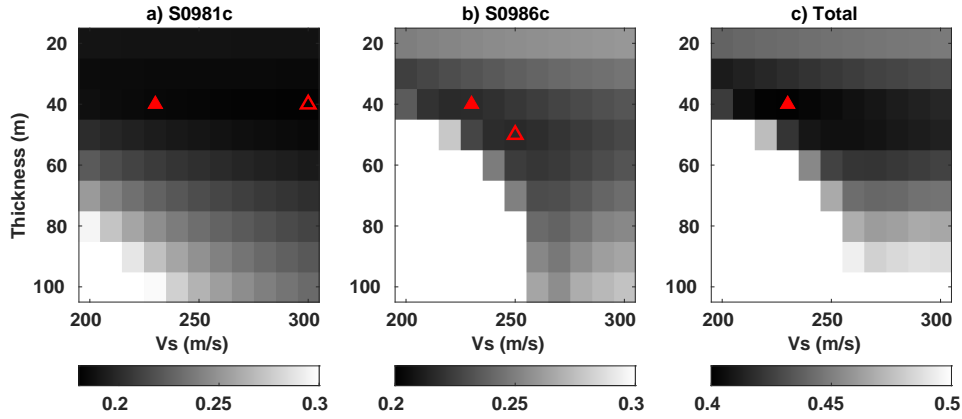


Figure 5. The misfit between the synthetic and observed chirp signals varies with the V_S and the thickness of the second subsurface layer (Table 1). The gray color indicates the misfit value. The red triangles correspond to the least misfit of the two events, while the empty triangle (a,b) corresponds to the least misfit of each event. The empty areas are the parameter combinations where the misfit values are larger than the color bar maximum or an air-coupled Rayleigh wave is excited.

5 Discussion

Our analytical waveform modelling aid us validating the Martian subsurface models from other observations (Section 5.1). By combining our modelling and the Martian atmospheric model, we provide an explanation for the S0793a seismic observation (Section 5.2). We also discuss how to improve our modelling in the future (Section 5.3 and 5.4).

5.1 Implication for the subsurface structure

Our waveform fitting provides a velocity model where the V_S increases with depth (Table 1). In this model, the first layer is interpreted as fine-sand-dominated regolith (e.g. Grott et al., 2021). Compared to the V_S of the first layer, the second-layer V_S is larger and thus corresponds well to a coarse regolith where pebbles exist. The second layer is of 40-m thickness, close to the estimate of the maximum surficial regolith thickness around InSight; however, the geological model beneath the InSight indicates that this coarse regolith layer is interrupted by a basalt layer (e.g. Warner et al., 2022). Note that the geological model is built by considering the velocity models from the horizontal-to-vertical (H/V) ratio measurement of the InSight ambient seismic vibration (Hobiger et al., 2021); we refer to the velocity models as the H/V models. To further investigate this layer thickness contradiction, we check how the synthetic waveforms from the H/V models fit the chirps.

We choose one of the H/V models (Figure 6a), which is close to the geological model, and then generate the synthetic waveforms (Figure 6b-e) following our computation in Section 4. We notice that for S0981c, the radial-component waveform of the H/V model possesses larger amplitudes than the observed chirp; for S0986c, the radial component from the H/V model does not fit the observed waveform around 5:29:50. The total misfit of the H/V model (Equation 22) is 0.44, 15% larger than the total misfit of our three-layer model, 0.40. Thus our model provides a better waveform fitting to the chirps than the H/V model. We repeat the above process with the other three H/V models (Figure S2,S3,S4). We notice that all the three models provides smaller radial-component amplitudes compared to the real chirps.

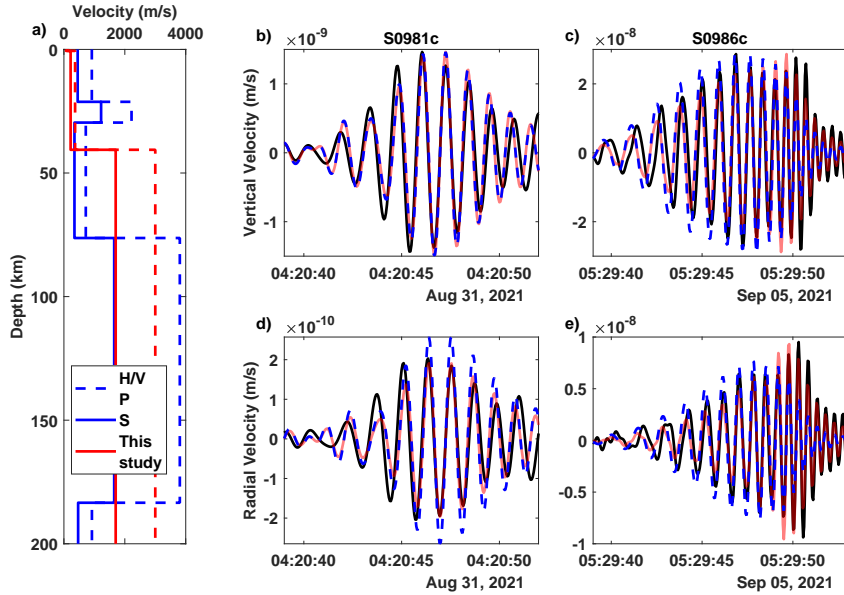


Figure 6. Comparison of the chirp waveform fitting between the H/V and the 3-layer models. The H/V model (blue) is the weakly-constrained maximum a posteriori estimation from Hobiger et al. (2021) and provides the blue dashed waveforms (b,c,d,e). The black and red waveforms (b,c,d,e) are the same observed and synthetic chirps from Figure 4, respectively.

388 The synthetic waveforms from the H/V models fit the vertical component of the
 389 recordings (Figure 6, S2, S3, S4). Thus the H/V models does not contradict with the chirp
 390 observation. Meanwhile, this waveform fitting difference between the H/V and our models
 391 could be due to the different sensitivities of the H/V ratio and the compliance to a
 392 same elastic property (e.g. Maupin, 2017; Kenda et al., 2020). Therefore, even through
 393 our model explain the chirps better than the H/V models, we cannot determine if our
 394 model is closer to the real subsurface than the H/V models. In order to achieve an ac-
 395 curate subsurface model, we need to incorporate the chirp observation, the H/V ratio
 396 measures, and other available data like the normalized compliance measures from the
 397 Martian pressure drops (e.g. Kenda et al., 2020; Onodera, 2022).

398 5.2 Higher-mode guided infrasound from the atmospheric model

399 The S0793a seismic recording includes a chirp signal in the time domain (Figure 7a).
 400 However, in opposition to the S0986c and S0981c group velocity measurement, which yields
 401 a single monotonous trend (Figure 3c and S1b), the measurement of the S0793a chirp
 402 provides two trends (Figure 7b): a horizontal one from 1 to 2 Hz at around 251 m/s, and
 403 a sloping one with decreasing velocity, from about 1.2 Hz to 2.3 Hz. The horizontal trend
 404 may correspond to an infrasound propagating directly from the impact and requires fur-
 405 ther investigation. It is ambiguous to attribute the sloping trend to a guided infrasound,
 406 since the bandwidth of the trend is short and the trend is not as continuous as the ones
 407 of S0981c and S0986c (Figure S1b and S1c). In the rest of this subsection, we present
 408 that the sloping trend can be explained by the higher-mode guided infrasound, while we
 409 also recognize that there could be other interpretations for the trend like the scattered
 410 infrasound (i.e. echoes, Garcia et al., 2022).

411 We compute the fundamental- and 1st-higher-mode group velocities and the eigen-
 412 functions of this guided infrasound by applying our computation (Section 2) to the cor-
 413 responding infrasound velocity model (Figure 7c). The synthetic fundamental-mode group

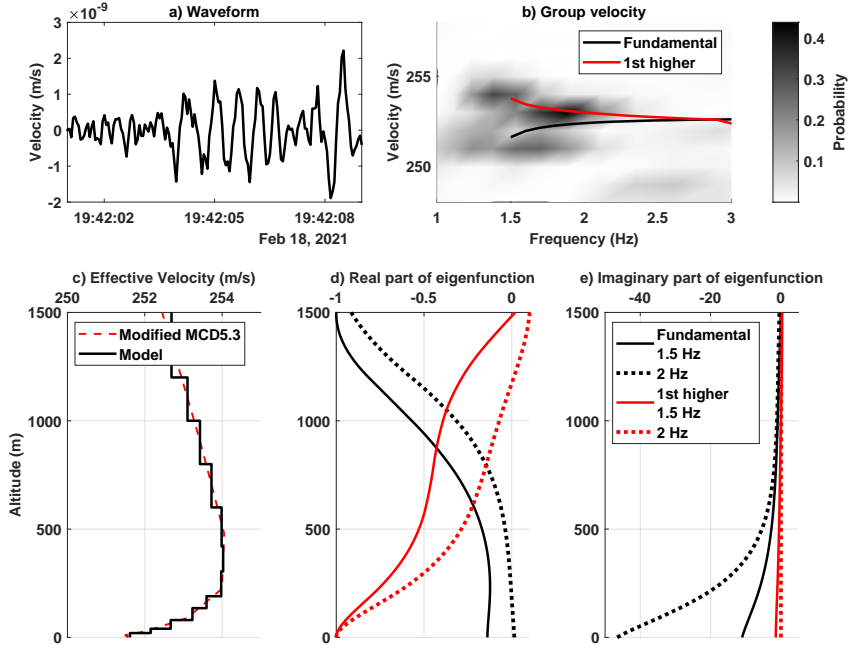


Figure 7. (a) The S0793a chirp bandpass filtered between 0.5 and 8 Hz. (b) Comparison of the group-velocity measurement from the chirp (gray) to the synthetic group velocity of the fundamental- and 1st-higher mode. These modes are from a 1D layered model modified from the MCD model (c, Table S1). We also compute the eigenfunctions corresponding to these modes and normalize each eigenfunction by its maximum real value (d and e).

414 velocity increases with increasing frequency, opposite of the measured group velocity (black
 415 curve in Figure 7b). The fundamental-mode eigenfunctions are complex valued, instead
 416 of real valued as in S0986c (Section 3b). The real parts of the eigenfunctions are close
 417 to zero on the ground surface and increase from the surface to the 1500-m altitude (Fig-
 418 ure 7d). This means that this fundamental-mode guided infrasound possesses smaller
 419 amplitude on the ground surface than at 1500 m altitude. The imaginary parts of the
 420 eigenfunctions are large near the ground surface (Figure 7e) and represent energy leak-
 421 age of the guided infrasound to the top halfspace (e.g. Press & Harkrider, 1962; Radovich
 422 & De Bremaecker, 1974). Thus this fundamental mode possesses weak pressure on the
 423 ground surface.

424 The 1st-higher-mode guided infrasound of the S0793a atmosphere model possesses
 425 a stronger pressure than the fundamental-mode on the ground surface (Figure S1d). We
 426 notice that the 1st-higher-mode group velocity agrees with the ground-velocity measure-
 427 ment (the sloping trend in Figure 7b). Furthermore, the real part of the 1st-higher-mode
 428 eigenfunctions presents a maximum absolute value on the ground surface, while the imag-
 429 inary part is close to zero on the ground surface compared to the fundamental-mode. There-
 430 fore, the 1st-higher-mode guided infrasound could generate the recorded S0793a chirp.

431 5.3 The free surface boundary condition

432 On the ground surface, the vertical particle velocity (v_z) in the atmosphere is equal
 433 to the vertical ground velocity (u_z). This yields:

$$434 \quad v_z(z = 0) = -u_z(z = 0), \quad (23)$$

436 where the minus sign is due to different conversion of the positive z -axis direction in at-
 437 mosphere (altitude) and the subsurface (depth). The formula above and Equation 4 lead
 438 to

$$439 \quad \partial_z p(z=0) = \rho(\partial_t + w_x \partial_x) u_z(z=0), \quad (24)$$

441 where ρ is the air density. We rewrite this equation as

$$442 \quad \partial_z P(z=0) = -i\rho(\omega - w_x k) \frac{U_z(z=0)}{T_{zz}(z=0)} P(z=0), \quad (25a)$$

$$443 \quad = -i\rho(\omega - w_x k) C_z P(z=0). \quad (25b)$$

445 The expected value for C_z on Mars is around $10^{-5} m/s/Pa$ and ρ is about $0.02 kg/m^3$.
 446 Thus, between 0.5 and 3 Hz, $\partial_z P(z=0)$ is on the order of 10^{-5} of $P(z=0)$. We can
 447 conclude that the guided-infrasound atmospheric pressure is almost constant near the
 448 ground surface and thus $v_z \approx 0$ (Equation 4). This justifies the rigid ground-surface
 449 boundary condition used in our derivation of guided infrasound (Equation 7). Our bench-
 450 mark (Appendix B) also validates that our modelling result agrees well with the numer-
 451 ical simulation where v_z is continuous on the ground surface, the atmosphere-solid-earth
 452 boundary.

453 We notice that the continuous- v_z boundary condition is necessary for modelling the
 454 coupled normal mode between a planet atmosphere and the solid planet (e.g. Watada,
 455 1995; Lognonné et al., 1998; Tanimoto, 2001). Thus in order to be able to model the cou-
 456 pled mode, we will incorporate this boundary condition into our modelling in future.

457 **5.4 Potential improvement and future work**

458 In our forward modelling, we assume the boundary between the atmosphere and
 459 the ground to be flat. However, the ground surface topography affects the compliance
 460 (e.g. Bishop et al., 2021) by altering the guided infrasound horizontal wavenumber rel-
 461 ative to the ground surface. To incorporate the topography into the compliance com-
 462 putation, we can compute the spatial wavenumber of the topography and combine the
 463 wavenumber with the one of guided infrasound, similar to the microseism studies which
 464 consider ocean waves coupling with topographic seafloors (e.g. Ardhuin et al., 2015).

465 Since our forward modelling is fast in computation, one can adopt this modelling
 466 to perform a Markov-chain Monte Carlo inversion of the atmospheric and subsurface pa-
 467 rameters from the chirp signals. A Markov-chain Monte Carlo inversion explores differ-
 468 ent parameter combinations (e.g. Tarantola, 2005). As we demonstrate the trade-off be-
 469 tween the shear-wave velocity and layer thickness in our three-layer subsurface model
 470 (Section 4), we expect more parameter trade-offs if the subsurface model possesses more
 471 than three layers. We also expect a trade-off between the atmospheric and subsurface
 472 parameters, since the compliance computation depends on the guided-infrasound phase
 473 velocity from the atmospheric model. Thus the inversion could aid us to assess these trade-
 474 offs between both the atmospheric (like infrasound velocity) and the subsurface (e.g. V_s
 475 and layer thickness) parameters.

476 **6 Conclusion**

477 We analytically model chirp, the seismic waveform due to the coupling between guided
 478 infrasound and the ground. We theoretically demonstrate the guided-infrasound prop-
 479 agation in a 1D atmospheric model and compute the guided-infrasound phase and group
 480 velocities. Our group velocities match well with the measures from the S0981c and S098c
 481 chirp observation on Mars. We convert the guided infrasound into a chirp through com-
 482 pliance of a three-layer subsurface velocity model. We validate our modelling through
 483 a benchmark.

484 By applying our modelling to the Martian atmospheric model, we model synthetic
 485 chirps to fit the real chirp recordings of S0981c and S0986c. Through the waveform fit-
 486 ting, we achieve a three-layer Martian subsurface velocity model, where the coarse re-
 487 golith is 40-m thick. We also apply our modelling to examine the velocity models from
 488 the InSight ambient-seismic-vibration H/V ratio observation and present that these mod-
 489 els explain a part of the chirp recordings but do not provide a good fitting as our three-
 490 layer model. Therefore we need to incorporate all the available observations to constrain
 491 the Martian subsurface structure estimation.

492 7 Open Research and Data Availability Statement

493 The Martian topography data are from the NASA PDS Geosciences Node (Neumann
 494 et al., 2003) The InSight seismic waveform data are available from the IGP Datacen-
 495 ter, IRIS-DMC and the NASA PDS (InSight Mars SEIS Data Service, 2019a, 2019b).
 496 The Martian atmospheric parameters are from MCD (Millour et al., 2018). The codes
 497 for computation of the guided infrasound are available in Xu (2022).

498 Appendix A Guided infrasound on Earth

499 We present a terrestrial example of chirp, where the infrasound source is the 2017-
 500 12-12 Baumgarten gas hub explosion in Austria. This event generates both seismic waves
 501 and infrasound. These infrasound couples to the ground and are recorded by local seis-
 502 mic stations (Schneider et al., 2018). From these stations, we choose one from AlpAr-
 503 ray (Hetényi et al., 2018), A333A (Figure A1), at the northeast of the explosion. The
 504 A333A vertical-component recording presents a chirp (Figure A1b), from which we mea-
 505 sure the group velocity (Figure A1d). Our group-velocity measurement method is de-
 506 tailed in Panning et al. (2015) and Drilleau et al. (2020). We choose an atmospheric pro-
 507 file used in Schneider et al. (2018) and follow the computation in Section 2 to compute
 508 the synthetic group velocity. The difference between the measured and synthetic group
 509 velocity indicates that we can improve the Earth low-altitude (<3000 m) atmospheric
 510 model using the guided infrasound.

511 Appendix B Benchmark

512 We benchmark our modelling against a numerical simulation software, SPEC-FEM2D-
 513 DG (Martire et al., 2020). SPEC-FEM2D-DG can model acoustic and seismic waves in
 514 a coupled solid-fluid system. Garcia et al. (2022) utilize this software to simulate the seis-
 515 mic recording at InSight location due to the S0793a, S0981c, and S0986c impacts. We
 516 choose the S0981c simulation result in this benchmark. Garcia et al. (2022) use the S0981c
 517 atmospheric model from MCD (Millour et al., 2018) and a four-layer subsurface model
 518 (Table B1). Garcia et al. (2022) set the pressure and seismic receivers 246 km away from
 519 the source and the seismic receiver at 5 m under the ground surface. We use the same
 520 models and the same receiver setting. We compare our modelling result to the one from
 521 Garcia et al. (2022): the group velocity (Figure B1a), the compliance (Figure B1b), and
 522 the pressure and seismic waveforms (Figure B2). Those match well in the frequency do-
 523 main or the time domain. Note that in the seismic waveform comparison, since SPEC-FEM2D-
 524 DG is for the 2D space, instead of the 3D, we have to modify Equation 20 and 20 by re-
 525 moving the 3D cylindrical wave term:

$$u_z = \mathcal{F}^{-1} \left[S(\omega) \exp(-ikx - ax) \frac{P^2(\omega, 0)}{\int P^2(\omega, z) dz} C_z(\omega) \right], \quad (\text{B1})$$

$$u_x = \mathcal{F}^{-1} \left[S(\omega) \exp(-ikx - ax) \frac{P^2(\omega, 0)}{\int P^2(\omega, z) dz} C_x(\omega) \right]. \quad (\text{B2})$$

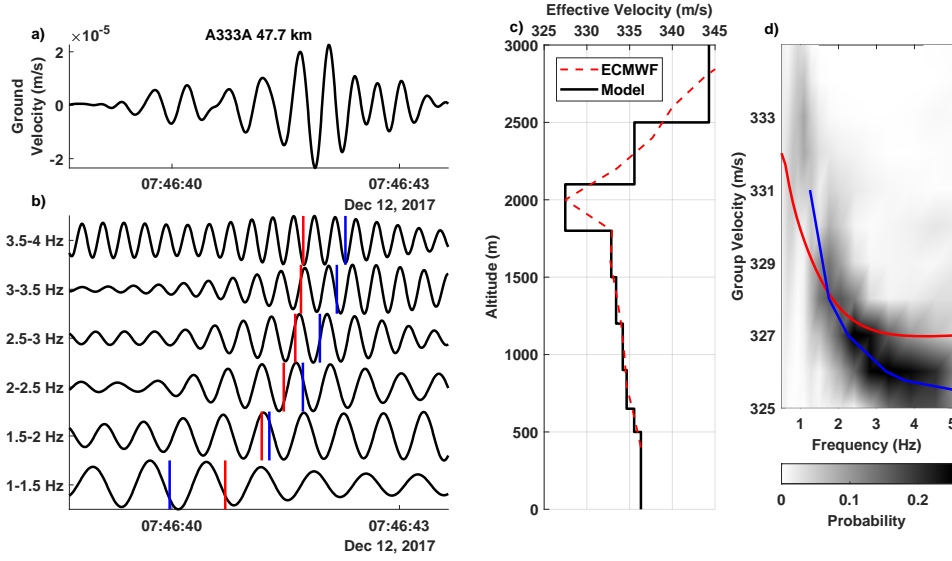


Figure A1. Illustration of the A333A chirp (a,b) and computation of the group velocity (c,d). We bandpass filter the A333A vertical-component velocity recording between 0.5 and 5 Hz (a) and in multiple narrow bands (b). We measure group velocity from the recording (a) and pick the probability maximum at each frequency as the group velocity (blue line in d). Based on the ECMWF profile used in Schneider et al. (2018), we plot the effective acoustic-wave velocity profile (red dashed line in c) and the 1D layered model (black line in c). From the 1D model, we compute the synthetic group velocity (red line in d). We convert the synthetic and measured group velocities at the narrow bands to the corresponding arrival times (red and blue bars in b).

Table B1. The subsurface velocity model used in (Garcia et al., 2022). Note that we only use this model in the benchmark section.

Layer number	$V_P(m/s)$	$V_S(m/s)$	Density (kg/m^3)	Thickness (m)
1	744	398	1800	100
2	3800	1850	2304	9900
3	4500	2800	2570	14000
4	6224	3753	2863	∞

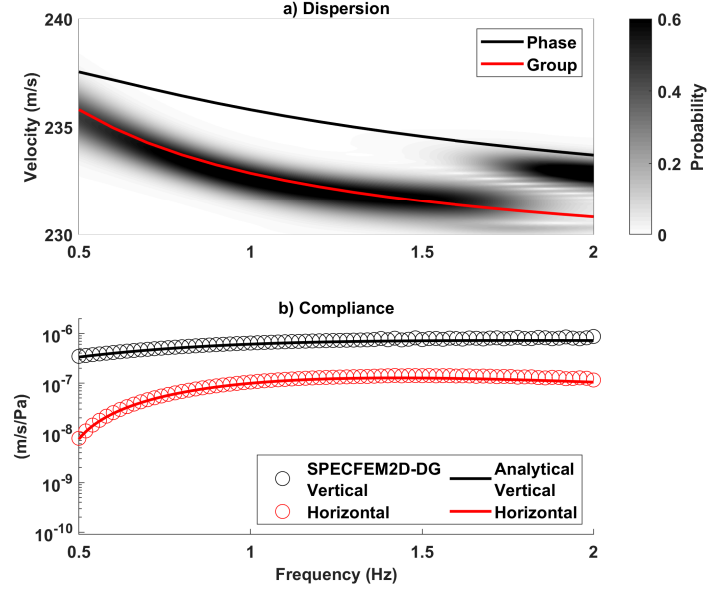


Figure B1. The comparison of the group velocity and compliance from our analytical approach to SPECFEM2D-DG. The group velocity measurement and compliance values from SPECFEM2D-DG are the gray background (a) and the empty circles (b), respectively. The estimates from our analytical approach are in the curves (a,b).

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Acknowledgments

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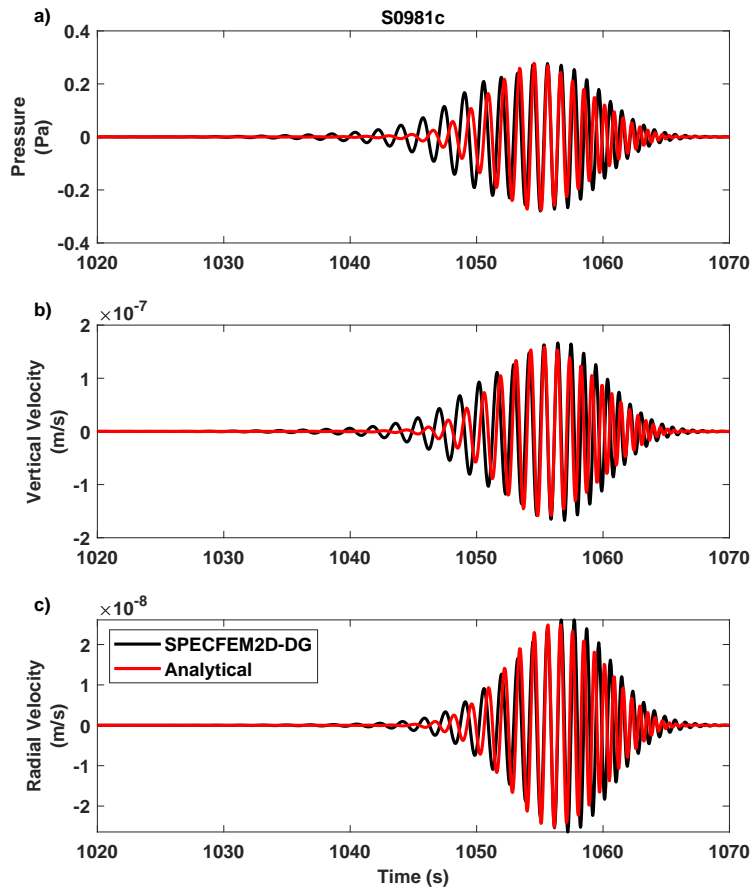


Figure B2. The waveform comparison of the atmospheric pressure, vertical velocity, and horizontal velocity from our analytical approach to SPECFEM2D-DG. The waveforms are all bandpass filtered between 0.5 and 2 Hz. We normalize our waveforms by the maximum value of the pressure from SPECFEM2D-DG.

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