

2D characterization of near-surface V_P/V_S : surface-wave profiling *versus* refraction tomography

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Abstract

The joint study of pressure (P-) and shear (S-) wave velocities (V_P and V_S), as well as their ratio (V_P/V_S), has been used for many years at large scales but remains marginal in near-surface applications. For these applications, V_P and V_S are generally retrieved with seismic refraction tomography combining P and SH (shear-horizontal) waves, thus requiring two separate acquisitions. Surface-wave prospecting methods are proposed here as an alternative to SH-wave tomography in order to retrieve pseudo-2D V_S sections from typical P-wave shot gathers and assess the applicability of combined P-wave refraction tomography and surface-wave dispersion analysis to estimate V_P/V_S ratio. We carried out a simultaneous P- and surface-wave survey on a well-characterized granite-micaschists contact at Plœmeur hydrological observatory (France), supplemented with an SH-wave acquisition along the same line in order to compare V_S results obtained from SH-wave refraction tomography and surface-wave profiling. Travel-time tomography was performed with P- and SH-wave first arrivals observed along the line to retrieve V_P^{tomo} and V_S^{tomo} models. Windowing and stacking techniques were then used to extract evenly spaced dispersion data from P-wave shot gathers along the line. Successive one-dimensional Monte Carlo inversions of these dispersion data were performed using fixed V_P values extracted from the V_P^{tomo} model and no lateral constraints between two adjacent one-dimensional inversions. The resulting one-dimensional V_S^{sw} models were then assembled to create a pseudo-2D V_S^{sw} section, which appears to be correctly matching the general features observed on the V_S^{tomo} section. If the V_S^{sw} pseudo-section is characterized by strong velocity uncertainties in the deepest layers, it provides a more detailed description of the lateral variations in the shallow layers. Theoretical dispersion curves were also computed along the line with both V_S^{tomo} and V_S^{sw} models. While the dispersion curves computed from V_S^{sw} models provide

results consistent with the coherent maxima observed on dispersion images, dispersion curves computed from V_S^{tomo} models are generally not fitting the observed propagation modes at low frequency. Surface-wave analysis could therefore improve V_S models both in terms of reliability and ability to describe lateral variations. Finally, we were able to compute V_P/V_S sections from both V_S^{sw} and V_S^{tomo} models. The two sections present similar features, but the section obtained from V_S^{sw} shows a higher lateral resolution and is consistent with the features observed on electrical resistivity tomography, thus validating our approach for retrieving V_P/V_S ratio from combined P-wave tomography and surface-wave profiling.

Keywords: Seismic methods, Surface-wave profiling, P-wave tomography, SH-wave tomography, V_P/V_S ratio.

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1. INTRODUCTION

The joint study of pressure (P-) and shear (S-) wave velocities (V_P and V_S , respectively), as well as their ratio (V_P/V_S), has been used for many years at large scales. V_P/V_S is commonly employed in seismology and geodynamics to study oceanic and continental crusts' structures (Nicholson and Simpson 1985; Julià and Mejía 2004; Tryggvason and Linde 2006; Powell *et al.* 2014), subduction and extension zones (Nakajima *et al.* 2001; Bauer *et al.* 2003; Latorre *et al.* 2004; Gautier *et al.* 2006; Reyners *et al.* 2006), active volcanic areas (Walck 1988; Sanders *et al.* 1995; Lees and Wu 2000; Schutt and Humphreys 2004), or earthquake-source regions (Catchings 1999; Ryberg *et al.* 2012). V_P/V_S has proved to be an efficient parameter to highlight the existence of melt or aqueous fluid phase (Takei 2002) since the liquid phase affects V_P and V_S differently (Biot 1956a,b).

Many theoretical studies (Berryman 1999; Lee 2002; Dvorkin 2008) and experimental developments (Wyllie, Gregory, and Gardner 1956; Murphy 1982; Prasad 2002; Uyanik 2011) have been aimed at understanding the effect of saturation and pore fluids on body wave velocities in consolidated media, especially in the field of hydrocarbon exploration where the V_P/V_S ratio is frequently used to discriminate different pore fluids in reservoirs (Tatham and Stoffa 1976; Fu, Sullivan, and Marfurt 2006; Rojas 2008). The value of the V_P/V_S ratio is also related to *in situ* stress orientation (Thompson and Evans 2000), fractures and cracks presence, and pore geometry for individual lithologies with small variations in composition (Tatham 1982; Wilkens, Simmons, and Caruso 1984).

In near-surface applications (at depth lower than 100 m), the combined study of V_P and V_S is often proposed without the calculation of V_P/V_S ratios. It is classically carried out for engineering purposes to determine the main mechanical properties of reworked materials in active landslides (Godio, Strobbia, and De Bacco 2006; Jongmans *et al.* 2009; Socco *et al.* 2010b; Hibert *et al.* 2012), control fill compaction in civil engineering (Heitor *et al.* 2012; Cardarelli, Cercato, and De Donno 2014), study weathering and alteration of bedrock (Olona *et al.* 2010), or assess earthquake site response (Jongmans 1992; Lai and Rix 1998; Raptakis *et al.* 2000; Othman 2005). More recently, this approach has also been proposed for hydrological applications to characterize shallow aquifers (Grelle and Guadagno 2009; Mota and Monteiro Santos 2010; Konstantaki *et al.* 2013; Pasquet *et al.* 2015).

For these shallow-target studies, V_P and V_S are generally retrieved with seismic refraction

31 tomography using both P and SH (shear-horizontal) waves (Turesson 2007; Grelle and Guadagno
32 2009; Fabien-Ouellet and Fortier 2014; Pasquet *et al.* 2015). The use of this method is widespread
33 since it is easily carried out with a one-dimensional (1D) to three-dimensional (3D) coverage, quick
34 to implement and relatively inexpensive (Galibert *et al.* 2014). However, if measurements of V_P
35 are performed quite efficiently for many years, retrieving V_S remains complex since it requires the
36 use of horizontal component geophones difficult to set up horizontally (Sambuelli *et al.* 2001) and
37 specific sources strenuous to use (Sheriff and Geldart 1995; Jongmans and Demanet 1993; Xia *et al.*
38 2002; Haines 2007).

39 As an alternative to SH-wave refraction tomography, surface-wave prospecting methods are
40 commonly proposed to achieve indirect estimation of V_S in a relatively straightforward manner
41 (e.g., Gabriels, Snieder, and Nolet (1987); Jongmans and Demanet (1993); Park, Miller, and Xia
42 (1999); Socco and Strobbia (2004); and Socco, Foti, and Boiero (2010a)). Due to their dispersive
43 nature, surface waves are characterized by an investigation depth that mainly depends on the
44 considered data frequency. Surface waves are thus widely used at large scales in global seismology for
45 mantle investigations using low frequencies. When targeting shallow structures with strong lateral
46 variability, surface-wave methods are, however, limited by the well-known trade-off between lateral
47 resolution and investigation depth (Gabriels *et al.* 1987). On the one hand, the inverse problem
48 formulation requires the investigated medium to be assumed 1D below the spread, which has to be
49 short enough to achieve lateral resolution and perform two-dimensional (2D) profiling. On the other
50 hand, long spreads and low-frequency geophones are required to record long wavelengths in order
51 to increase the investigation depth and mitigate near-field effects (Russel 1987; Forbriger 2003a,b;
52 O’Neill 2003; O’Neill and Matsuoka 2005; Bodet *et al.* 2005; Zywicki and Rix 2005; Bodet, Abraham,
53 and Clorennec 2009). When the seismic set-up provides redundant data, several countermeasures
54 exist to overcome these drawbacks and narrow down the lateral extent of dispersion measurements,
55 such as common mid-point cross-correlation (Hayashi and Suzuki 2004; Grandjean and Bitri 2006;
56 Ikeda, Tsuji, and Matsuoka 2013), multi-offset phase analysis (Strobbia and Foti 2006; Vignoli and
57 Cassiani 2010), or offset moving windows and dispersion stacking techniques (O’Neill, Dentith, and
58 List 2003; Bohlen *et al.* 2004; Neducza 2007; Boiero and Socco 2010, 2011; Bergamo, Boiero, and
59 Socco 2012).

60 The joint analysis of travel-time tomography V_P and surface-wave profiling V_S has recently

61 been proposed to retrieve 1D time-lapse V_P/V_S soundings (Pasquet *et al.* 2015) or 2D V_P/V_S
62 sections (Ivanov *et al.* 2006; Konstantaki *et al.* 2013). Pasquet *et al.* (2015) highlighted an overall
63 consistency between the temporal variations of the water table and V_P/V_S contrasts. For their
64 part, Konstantaki *et al.* (2013) assessed the lateral fluctuations of a shallow aquifer water table
65 level with 2D V_P/V_S variations. Using a single standard acquisition set-up to retrieve 2D V_P and
66 V_S sections thus appears interesting and convenient to reduce equipment costs and acquisition
67 time. Yet, refraction tomography and surface-wave profiling involve distinct characteristics of the
68 wavefield and different assumptions about the medium, thus providing results of different resolutions
69 and investigation depths difficult to compare to each other.

70 This study tackles such issues through a systematic comparison of V_S models obtained from
71 SH-wave refraction and surface-wave dispersion inversion, along with V_P retrieved from P-wave
72 refraction, as recently proposed by Pasquet *et al.* (2014, 2015). For this purpose, we targeted
73 the Plœmeur hydrological observatory (France). This experimental site has been subject to many
74 geophysical and hydrogeological studies aimed at characterizing the flow processes involved in the
75 recharge of the outstandingly productive fractured aquifer present in the region (Touchard 1999;
76 Le Borgne *et al.* 2006a,b, 2007; Ruelleu *et al.* 2010; Jiménez-Martínez *et al.* 2013). The study
77 area is located at a contact between granites and micaschists, clearly highlighted in the surface by
78 electrical resistivity tomography (ERT) and electrical conductivity (EC) mapping. However, previous
79 refraction seismic studies showed that V_P alone was neither able to detect the contact zone nor able
80 to discriminate granites from micaschists, probably because P-wave velocity is mainly controlled by
81 the water content in the weathered areas. The site consequently provided a challenging framework
82 to test the applicability of the joint interpretation of V_P and V_S for near-surface applications.
83 In the present study, V_P and V_S sections were classically obtained with P- and SH-wave travel-
84 time tomography carried out on a line intersecting the contact zone. Surface-wave profiling was
85 performed by means of offset moving window and dispersion stacking techniques. Local dispersion
86 measurements were first extracted from different shots illuminating the same portion of the seismic
87 line and then stacked to increase signal-to-noise ratio. The extraction window was eventually moved
88 along the line to retrieve a collection of local multimodal dispersion measurements. Several window
89 lengths were tested to find the best compromise between lateral resolution and investigation depth
90 (Pasquet *et al.* 2012). The lateral consistency of dispersion data was thoroughly controlled during

91 picking through visual browsing and *a posteriori* verified on phase velocity pseudo-sections. Separate
92 Monte Carlo inversions of dispersion curves were then performed along the line with no lateral
93 constraints in order to reconstruct a pseudo-2D V_S section. The parameterization of those inversions
94 was based on: (i) V_P obtained from travel-time tomography; (ii) *a priori* geological knowledge;
95 and (iii) maximum wavelengths observed along the line. Theoretical dispersion curves were then
96 recomputed from both V_S models along the line to control the inversion quality and the consistency
97 of these models. Finally, V_P/V_S obtained from both methods were compared to evaluate their ability
98 to image V_P/V_S variations and assess their practical limitations.

99 2. SITE DESCRIPTION AND DATA ACQUISITION

100 2.1. Geological setting

101 The Plœmeur site is located on the south coast of Brittany (west of France), 3 km far from the
102 Atlantic Ocean, near the city of Lorient (Fig. 1). The crystalline bedrock aquifer present in the area
103 is composed of tectonic units developed during the Hercynian orogeny and marked by numerous
104 synkinematic intrusions of upper Carboniferous leucogranites (Ruelleu *et al.* 2010). The pumping
105 site is located at the intersection of: (i) a contact between the Plœmeur granite and overlying
106 “Pouldu” micaschists dipping 30° to the North and (ii) a subvertical fault zone striking N 20° (Fig. 1).
107 Weathering in the area is limited to the first few metres, except in the micaschists near the pumping
108 site where it reaches about 30 m. Before the start of the pumping activities in 1991, the site was a
109 natural aquifer discharge area with preferential upward fluxes. The average water levels began to
110 decline during the first years of operation but have stabilized since 1997 (Jiménez-Martínez *et al.*
111 2013). Despite the low permeability and porosity of these lithologies, pumping wells implanted in
112 the site have been continuously producing water at a rate of about 10^6 m³ per year since 1991
113 (Touchard 1999), with limited head decrease and no seawater intrusion. One of the challenges on
114 this site is to understand recharge processes in these highly heterogeneous systems. For this purpose,
115 the site is monitored by several wells implanted mostly around the contact zone and in the clayey
116 area overlaying micaschists (F* and MF* in Fig. 2).

117 2.2. Previous geophysical results

118 Slingram EC mapping and ERT were carried out on the site prior to the seismic campaign in
119 order to accurately describe near-surface lithologies. Apparent EC (σ_a) variations over the first

120 5.5 m in depth were mapped using an electromagnetic device with low induction number (intercoil
121 spacing of 3.66 m and frequency of 9.8 kHz) in vertical dipole (VD) configuration, integrating
122 conductivity values down to about 6 m in depth (McNeill 1980). We used a continuous acquisition
123 mode following profiles separated with 5 m to 7 m, covering an area of about 15 ha (Fig. 2a). As for
124 ERT, we used a multi-channel resistivimeter with a 96-electrode Wenner–Schlumberger array and 1
125 roll-along (Fig. 2b). The electrodes were spaced with 4 m in order to obtain a 476-m-long profile
126 roughly oriented west–east (WE on Fig. 2a). The inversion was performed using the RES2DINV
127 software (Loke and Barker 1996).

128 Results of EC mapping show smooth lateral variations of σ_a (from less than 5 mS/m to over
129 30 mS/m, i.e., from 200 Ωm to less than 30 Ωm in terms of apparent electrical resistivity ρ_a) in
130 the subsurface. Western low σ_a values are clearly associated with the presence of very shallow
131 granite (between 0 m and 125 m along the WE profile in Fig. 2a). On the contrary, higher σ_a values
132 observed in the eastern part can be related to clays overlaying weathered micaschists (between 275 m
133 and 476 m in Fig. 2a). Such distribution seems in agreement with the assumption of the contact
134 zone striking N 20° in the area. ERT results are also consistent with the anticipated geological
135 structures in depth and clearly match the apparent EC variations in surface (Fig. 2b). Four main
136 structures can be delineated in Fig. 2b: fresh granite (FG), almost outcropping in the western part,
137 characterized by high-electrical-resistivity (ρ) values (around 1000 Ωm); weathered granite (WG),
138 at the surface in the western part, characterized by significantly lower ρ values (around 200 Ωm);
139 clays (CL), at the surface in the eastern part, characterized by slightly lower ρ values (between
140 50 Ωm and 200 Ωm); and micaschists (MS), deeper in the eastern part, characterized by higher ρ
141 values (around 750 Ωm). Possible evidence of the contact zone is visible between 225 m and 250 m
142 along the ERT profile, marking a strong contrast between WG and clays (hashed area in Fig. 2b).
143 The positions of the nearest piezometric wells are projected on the WE line with their measured
144 piezometric head level. A significant decrease in this level is observed in the wells located in the
145 hashed area due to pumping occurring in the F28 well. While Pasquet *et al.* (2015) studied V_P/V_S
146 in simplified 1D conditions, this site offers the perfect framework to address 2D issues (i.e., presence
147 of strong lateral variations of lithology and topography), especially when knowing that previous
148 P-wave tomography studies failed to depict these lateral variations.

149 *2.3. Seismic acquisition*

150 The seismic acquisition set-up was deployed along the ERT profile (WE in Fig. 2a). It consisted
151 of a simultaneous P- and surface-wave acquisition followed by an SH-wave acquisition along the same
152 line. We used 72 geophones, a 4-m receiver spacing and 2 roll-alongs to finally obtain a 476-m-long
153 profile (Fig. 3). A 72-channel seismic recorder was used with 14-Hz vertical component geophones
154 for the P-wave and surface-wave profiles and with 14-Hz horizontal component geophones for the
155 SH-wave profile. The use of 14-Hz geophones offers a good compromise to obtain seismic records
156 with a frequency content suitable for surface-wave, refraction, and reflection processing. First shot
157 location was one-half receiver spacing away from the first trace, and move-up between shots was one
158 receiver interval in order to achieve the high coverage required to perform refraction tomography
159 and to stack dispersion data. In addition, 219 shots were recorded along each profile for a total
160 number of 15768 active traces. The P-wave source consisted of an aluminium plate hit vertically by
161 a 5-kg sledgehammer. The plate was hit six times at each position to increase signal-to-noise ratio.
162 SH waves were generated with a handheld source consisting of a heavy metal frame hit laterally
163 by a 5-kg sledgehammer. The SH-wave source was hit 12 times at each position. For both P- and
164 SH-wave acquisitions, the sampling rate was 0.25 ms and the recording time was 2 seconds to
165 include full surface-wave wavefield. A delay of -0.02 seconds was kept before the beginning of each
166 record to prevent early triggering issues. The collected data presented on Fig. 4 are affected with a
167 significant noise level, especially at far offset (80 m and more), due to active pumping wells and
168 military airplanes regularly landing and taking off from a nearby air force base. Seismograms clearly
169 show lateral variations due to both topographic effects and subsurface velocity changes, with strong
170 attenuation over clays and micaschists in the eastern part of the line.

171 **3. TRAVEL-TIME TOMOGRAPHY**

172 *3.1. Travel-time data*

173 We applied DC removal and a zero-phase low-pass filter on both P- and SH-wave data to remove
174 high-frequency noise (>100 Hz) and help for first arrival identification at far offsets. For the P-wave
175 profile, a total of 7076 first arrivals (45% of all traces) were determined (Fig. 5a) and 10352 traces
176 (65% of all traces) were picked on SH-wave shots (Fig. 5c). Travel-time data could not be obtained
177 where noise level was too high (grey dots in Fig. 5), with average maximum offsets around 100 m

178 for P waves and around 150 m for SH waves. Travel times are represented in a source–receiver
179 diagram (Fig. 5), which shows the distribution of first arrival times for each source–receiver pair
180 (Bauer *et al.* 2010; Baumann-Wilke *et al.* 2012). The diagonal traces indicate the zero-offset travel
181 times, where source and receiver locations are identical. This representation gives a first impression
182 of the subsurface velocity structure and allows for checking the lateral consistency of picked travel
183 times. Three main areas are clearly visible on both P- and SH-wave first arrivals: the first area
184 (from 0 m to 125 m) is characterized by the shortest first arrival times (i.e., shallow high velocity
185 zone); the area between 125 m and 275 m depicts slightly increasing times; and the third area
186 (from 275 m up to the end of the line) shows greater arrival times (compared with the first area),
187 probably associated with shallow low velocities.

188 The distribution depicted in Fig. 5a and 5c appears interestingly related to the lateral variability
189 observed on the EC map (Fig. 2a). Along the seismic profile, very low σ_a values occur between
190 0 m and 125 m on the granite side and are coherent with short first arrival times. High- σ_a values
191 observed between 275 m and the end of the line are also consistent with the shallow low velocity
192 zone suggested by long first arrival times. In the centre part (from 125 m to 275 m), intermediate
193 apparent EC values are in agreement with slightly decreasing shallow velocities observed in Fig. 5a
194 and 5c, suggesting a thickening of the weathered layer.

195 3.2. Tomography inversion

196 Travel-time data were inverted with the refraction tomography software RAYFRAC (Schuster
197 and Quintus-Bosz 1993; Sheehan, Doll, and Mandell 2005; Rohdewald 2011) using a smooth gradient
198 initial model. This model is the 2D extension of the mean 1D model obtained directly from picked
199 travel times, assuming velocity gradients in a 1D tabular medium (Gibson, Odegard, and Sutton
200 1979). It varied from 10 m/s at the surface to 3500 m/s at 40 m in depth for the P-wave model and
201 from 10 m/s at the surface to 2000 m/s at 40 m in depth for the SH-wave model. The inversion
202 process was stopped when velocity update, global root-mean-square (RMS) error, and maximum
203 normalized residual reached minimum values. In addition, 30 iterations were needed for both P-
204 and SH-wave travel times. For the P-wave profile, the global RMS error (in blue in Fig. 6b) and the
205 maximum normalized residual (in red in Fig. 6b) both steeply decrease during the first six iterations,
206 i.e., from 5 ms to 2.5 ms and from 25 ms to 13 ms, respectively, and more gradually during the last
207 iteration steps until they both reach minimum values of 2.1 ms and 11.5 ms, respectively. Mean

208 velocity update also rapidly decreases to reach a minimum value after six iterations (in black in
209 Fig. 6c). Maximum and minimum velocity updates (in red and blue, respectively, in Fig. 6c) are
210 more perturbed and tend to stabilize only after 20 iterations. As for SH-wave data, the global
211 RMS error (in blue in Fig. 6e) and the maximum normalized residual (in red in Fig. 6e) both
212 quickly decrease during the first 10 iterations, i.e., from 12.5 ms to 6 ms and from 50 ms to 22 ms,
213 respectively, and more smoothly during the last iteration steps until they both reach minimum
214 values of 4.3 ms and 20.1 ms, respectively. Mean, maximum, and minimum velocity updates (in
215 black, red, and blue, respectively, in Fig. 6f) reach minimum values after 10 iterations.

216 Normalized residuals were computed between observed and calculated travel times for every
217 picked trace (Bauer *et al.* 2010). Their distributions are represented in source–receiver diagrams for
218 both P- (Fig. 5b) and SH-wave (Fig. 5d) travel times. P-wave inversion results have lower global
219 RMS values (2.1 ms for P waves; 4.3 ms for SH waves) and maximum residual values (11.5 ms for
220 P waves; 20.1 ms for SH waves), and the mean residual value is higher for P waves (6.7%) than for
221 SH waves (3.9%). The ratio of residuals below 10% is also higher for SH-wave models (around 93%)
222 than for P-wave models (around 91%). The distribution of residuals shows the highest values at near
223 offsets, probably due to shallow lateral heterogeneities and triggering issues during the acquisition.

224 The final velocity models were clipped for both V_P (Fig. 6a) and V_S (Fig. 6d) sections below a
225 ray coverage of 100 rays per grid cell to keep only well-resolved areas. Both models are characterized
226 by velocities mainly following a linearly increasing trend in depth and show no strong lateral
227 variations. Only slight perturbations of the initial gradient model along the line can be observed
228 with: (i) higher V_P and V_S in the western part (HV_P and HV_S , respectively), corresponding to the
229 high-electrical-resistivity values associated with FG; (ii) lower V_P in the centre (LV_P), consistent
230 with a decrease in the piezometric head level and evidences of the contact zone inferred from ERT;
231 and (iii) lower V_S in the eastern part of the profile (LV_S).

232 4. SURFACE-WAVE PROFILING

233 4.1. Extraction of dispersion curves

234 Our multifold acquisition set-up allowed us to obtain surface-wave dispersion images from
235 P-wave shot gathers using windowing and dispersion stacking to narrow down the lateral extent of
236 dispersion measurements and increase the signal-to-noise ratio. Dispersion stacking was performed

237 following the basic workflow derived from O'Neill *et al.* (2003).

- 238 1. Select nW traces centred on a specific position along the line ($Xmid$).
- 239 2. Load a shot illuminating the selected spread.
- 240 3. Window the nW traces in the selected shot record.
- 241 4. Transform the wavefield to the frequency-phase velocity domain (dispersion image).
- 242 5. Normalize amplitude spectrum at each frequency.
- 243 6. Repeat steps 2–5 for the nS selected shots.
- 244 7. Stack all normalized dispersion images and pick the dispersion curves for each identified propa-
245 gation mode from the final image.
- 246 8. Shift the window of dW traces to the next $Xmid$ and repeat steps 1–7.

247 As aforementioned, the difficulty is to find the best compromise between investigation depth,
248 spectral resolution, and lateral resolution while keeping the 1D assumption valid. Thus, we first
249 performed trial-and-error tests (Pasquet *et al.* 2012) to select the optimum window size (nW)
250 and the number of sources (nS), keeping in mind that there is not, for the moment, any perfect
251 criterion (Pérez Solano 2013; Pérez Solano, Donno, and Chauris 2014). A 20-m-wide window with
252 six traces ($nW = 6$) was eventually used with six direct and six reverse shots on each side of the
253 window ($nS = 6$). Using more sources is likely to increase signal-to-noise ratio and enhance the
254 maxima but would narrow down the effective study area along the line (black dots in Fig. 2a).
255 Single dispersion images were obtained from each shot using a slant stack in the frequency domain
256 (Russel 1987; Mokhtar, Herrmann, and Russell 1988). These images (in which maxima should
257 correspond to Rayleigh-wave propagation modes) were first compared to confirm the validity of
258 the 1D approximation below the spread (Jongmans *et al.* 2009). These 12 single dispersion images
259 were then stacked as a final dispersion image. The moving window was finally shifted along the line
260 with a step of one receiver spacing ($dW = 1$, i.e., 4 m). We thus obtained evenly spaced dispersion
261 images at each spread mid-point ($Xmid$) with a large overlap in order to retrieve smoothly varying
262 dispersion images between two adjacent stacking windows and help for visual browsing when picking
263 dispersion curves.

264 We eventually obtained a collection of 105 stacked dispersion images along the line, on which
265 coherent maxima associated with the different propagation modes were identified. Visualization
266 of adjacent stacked dispersion images allowed for following the progressive lateral evolution of

267 the different modes and for avoiding mode misinterpretation (Zhang and Chan 2003; O’Neill and
 268 Matsuoka 2005; Boaga *et al.* 2013; Ezersky *et al.* 2013). These maxima were finally extracted on
 269 each stacked dispersion image with an estimated standard error in phase velocity defined according
 270 to the workflow described in O’Neill (2003). The dispersion lateral variability is illustrated here
 271 by two examples at both sides of the line (Fig. 7a ($X_{mid} = 50$ m) and 7c ($X_{mid} = 402$ m)).
 272 As, for instance, recommended by Bodet (2005) and Bodet *et al.* (2009), dispersion curves were
 273 limited down to frequencies (f_{lim}) where the spectral amplitude of the seismogram became too low
 274 (13.5 Hz in Fig. 7b and 8 Hz in Fig. 7d). Several authors (e.g., O’Neill (2003) and Zywicki and Rix
 275 (2005)) also mentioned that wavelengths higher than 50% of the spread length should not be used
 276 in order to mitigate near-field effects and prevent from velocity underestimation at low frequency.
 277 These recommendations are only basic rules of thumb mostly valid when using the fundamental
 278 mode only. Here, we used wavelengths higher than 50% of our spread length since we were able to
 279 perform dispersion stacking and use higher modes, which are of great constraint on the inversion
 280 with a strong impact on the investigation depth (Gabriels *et al.* 1987; Xia *et al.* 2003). Furthermore,
 281 possible low-frequency discrepancies were limited by attributing important errors to dispersion data
 282 with respect to frequency and spread length (O’Neill 2003). Finally, the corresponding maximum
 283 wavelength (λ_{max}) was extracted (35 m in Fig. 7a and 37 m in Fig. 7c) to retrieve $\lambda_{max}/2$, a typical
 284 investigation depth criterion (O’Neill 2003).

285 Up to four propagation modes were observed along the line and identified as fundamental (0),
 286 first (1), second (2), and third (3) higher modes (only modes up to 2 were identified in the examples
 287 shown in Fig. 7). The resulting dispersion curves are presented in Rayleigh-wave phase velocity
 288 pseudo-sections as a function of the wavelength λ and the spread mid-point X_{mid} (Fig. 8) in order
 289 to control the lateral coherence of mode identification (Strobbia *et al.* 2011; Haney and Douma 2012;
 290 Boiero *et al.* 2013a; Ezersky *et al.* 2013). The fundamental mode pseudo-section (Fig. 8a) does not
 291 present unrealistic abrupt changes (considering the overlap between two adjacent stacking windows)
 292 but shows significant lateral variations of Rayleigh-wave phase velocities. High phase velocities
 293 (from 200 m/s to 550 m/s) exist in the western part of the line (from the beginning to around
 294 100 m), whereas the eastern part of the line (from 275 m to the end) is characterized by lower
 295 phase velocities (around 150 m/s–200 m/s). The first (Fig. 8b), second (Fig. 8c), and third (Fig. 8d)
 296 higher modes show less lateral variations in terms of velocity, but the available frequency range for

297 mode 1 presents significant lateral fluctuations. The maximum wavelength (λ_{max}) observed at each
 298 X_{mid} ranges from 20 m to 50 m with an average value around 30 m (Fig. 8e).

299 4.2. Inversion

300 Assuming a 1D tabular medium below each extraction window, we performed a 1D inversion of
 301 dispersion data obtained at each X_{mid} . We used the neighbourhood algorithm (NA) developed
 302 by Sambridge (1999) and implemented for near-surface applications by Wathelet, Jongmans, and
 303 Ohrnberger (2004) and Wathelet (2008) within the GEOPSY tool (available at www.geopsy.org).
 304 Theoretical dispersion curves were computed from the elastic parameters using the Thomson–Haskell
 305 matrix propagator technique (Thomson 1950; Haskell 1953). NA performs a stochastic search of
 306 a pre-defined parameter space (namely, V_P , V_S , density, and thickness of each layer) using the
 307 following misfit function (MF):

$$MF = \sqrt{\sum_{i=1}^{N_f} \frac{(V_{cal_i} - V_{obs_i})^2}{N_f \sigma_i^2}}, \quad (1)$$

308 with V_{cal_i} and V_{obs_i} being the the calculated and observed phase velocities at each frequency f_i , N_f
 309 being the number of frequency samples, and σ_i being the phase velocity measurement error at each
 310 frequency f_i .

311 An appropriate choice of these parameters is considered as a fundamental issue for the successful
 312 application of inversion (Socco and Strobbia 2004; Renalier *et al.* 2010). Based on site *a priori*
 313 geological knowledge (presence of weathering gradients), we used parameterization with a stack
 314 of ten layers overlaying the half-space to look for a velocity gradient. The thickness of each layer
 315 was allowed for ranging from 0.5 m to 2.5 m. The maximum half-space depth (HSD) is of great
 316 importance since it depends on the poorly known investigation depth of the method. It was fixed to
 317 the half of the maximum wavelength observed along the entire line (25 m), as recommended by
 318 O’Neill (2003) and Bodet *et al.* (2005). The valid parameter range for sampling velocity models
 319 was 10 m/s–1500 m/s for V_S (based on dispersion observations and refraction tomography), with
 320 velocities constrained to only increase with depth, based on geological *a priori* information. P-wave
 321 velocity being of weak constraint on surface-wave dispersion, only the S-wave velocity profile can be
 322 interpreted (Der and Landisman 1972; Russel 1987). However, an identical layering is required for
 323 V_P and V_S in order to interpret V_P/V_S ratios. For this purpose, we extracted an average V_P value

324 for each 2.5-m-thick slice of the V_P model obtained from refraction tomography (Fig. 6a). This
 325 average value was then used to fix V_P in each layer of the inversion parameterization. Furthermore,
 326 V_S values were allowed to vary in such a way that Poisson's ratio values always remained between
 327 0.1 and 0.5 in order to prevent from unrealistic V_S values. Density was set as uniform (1800 kg/m^3)
 328 since its influence on dispersion curves is very limited (Der and Landisman 1972; Russel 1987). It is
 329 worth mentioning that, except for the V_P values, we used the same parameterization for all the 1D
 330 inversions performed along the line. We assumed that stacking and windowing already naturally
 331 smoothed the dispersion data, thus not requiring the use of lateral constraints between successive
 332 inversions.

333 A total of 63000 models were generated with NA (Fig. 9a ($X_{mid} = 50 \text{ m}$) and 9c ($X_{mid} =$
 334 402 m)). Models matching the observed data within the error bars were selected, as suggested
 335 by Endrun *et al.* (2008). The accepted models were used to build a final average velocity model
 336 associated with the centre of the extraction window (dashed line in Fig. 9b ($X_{mid} = 50 \text{ m}$) and 9d
 337 ($X_{mid} = 402 \text{ m}$)). As the maximum HSD remains constant along the line (same parameterization
 338 for each inversion), $\lambda_{max}/2$ is given (solid black line in Fig. 9b and 9d) to show where inverted
 339 models expand below typical investigation depth criterion. Normalized residuals between observed
 340 and calculated phase velocities were computed along the line for each individual sample of the picked
 341 dispersion curves. Their distributions are represented in pseudo-sections to control the quality of
 342 the final pseudo-2D V_S section. The fundamental mode residuals' pseudo-section (Fig. 10a) shows
 343 quite uniform values, with a maximum of 19%, a mean residual of 5%, and 86% of the residuals
 344 with values below 10%. Residual values obtained for the first higher mode (Fig. 10b) present higher
 345 values, especially at great wavelength. The maximum residual value is 29%, the mean is 9%, and
 346 only 62% of the residuals are below 10%. As for the second and the third higher modes (Fig. 10c
 347 and 10d, respectively), residuals remain very low with a maximum of 12% and 3%, respectively, and
 348 a mean residual of 1.8% and 1.3%, respectively. In addition, 99% of the second higher mode samples
 349 have residual values below 10%, whereas all samples of the third higher mode have residuals below
 350 10%. We additionally computed the misfit for each 1D V_S model along the line with equation (1)
 351 (Fig. 10e). Misfit values remain stable along the line and range from 0.05 to around 0.25, with a
 352 mean value of about 0.125. Several gaps are present along the line and correspond to inversions
 353 where none of the calculated models were fitting the error bars.

354 Each 1D V_S model was then represented at its corresponding X_{mid} position to obtain a pseudo-
 355 2D V_S section (Fig. 11). All the models were represented down to the maximum HSD (25 m),
 356 with the investigation depth criterion $\lambda_{max}/2$ superimposed in hashed black line. With such a
 357 representation, the actual HSD of each model can be easily followed along the line and compared
 358 with the investigation depth criterion. If the lateral variations of V_S values remain remarkably
 359 smooth in the shallow layers, the deepest layers and the half-space present an important variability
 360 of V_S caused by the higher uncertainties in dispersion measurements at great wavelength (i.e.,
 361 higher residual values in Fig. 10). Global results show a shallow low velocity layer (~ 250 m/s),
 362 which is thinner in the western part of the line (from 3 m to 6 m) and becomes thicker (~ 10 m) in
 363 the eastern part. High velocities (between 500 m/s and 1000 m/s) can be observed in the western
 364 part, directly below the shallow low velocity layer, whereas the velocity of the half-space remains
 365 below 500 m/s in the eastern part.

366 **5. QUALITATIVE COMPARISON OF VELOCITY MODELS AND RESULTING** 367 **V_P/V_S**

368 The comparison of velocity models obtained from P-wave tomography (V_P^{tomo} , Fig. 6a), SH-wave
 369 tomography (V_S^{tomo} , Fig. 6b), and surface-wave dispersion profiling (V_S^{sw} , Fig. 11) provided results
 370 that are consistent with the main structures interpreted from ERT data (Fig. 2b). However, velocity
 371 models do not provide such a clear delineation of these structures, especially for V_P^{tomo} and V_S^{tomo}
 372 sections. Indeed, the travel-time tomography method smoothes the lateral variations of velocity and
 373 often suffers from the strong influence of triggering issues on short travel times, which mainly affect
 374 the reconstruction of shallow structure velocities. As for surface-wave profiling, the overlaying of
 375 structures delimited by ERT on the V_S^{sw} pseudo-2D section (Fig. 11) confirms the lateral consistency
 376 of both V_S models in the first 20 m in depth. If the surface-wave method is clearly limited by its
 377 low investigation depth in this case, it provides more information regarding the lateral variations of
 378 shallow layers' velocities and seems to detect the modifications of mechanical properties occurring
 379 in the contact zone.

380 At both $X_{mid} = 50$ m (Fig. 12a) and $X_{mid} = 402$ m (Fig. 12d), 1D models of V_S and V_P
 381 extracted from tomography sections (V_S^{tomo} in green solid line and V_P^{tomo} in green dashed line,
 382 respectively) are characterized by similar trends of continuously increasing velocities in depth.

383 Furthermore, 1D V_S^{sw} models (red solid line) show the presence of two constant velocity layers
 384 followed by a linearly increasing velocity layer overlaying the half-space. Despite a low investigation
 385 depth, the V_S^{sw} pseudo-section manages to depict the shallow lateral variations and remains in good
 386 agreement with V_S^{tomo} .

387 As a control of both V_S models, forward modelling was performed along the line (examples
 388 for $X_{mid} = 50$ m and $X_{mid} = 402$ m are shown in Fig. 12c and 12f, respectively). On the one
 389 hand, theoretical dispersion curves were computed using 1D V_P^{tomo} and V_S^{tomo} models (green solid
 390 line). On the other hand, theoretical dispersion curves were calculated using 1D V_S^{sw} models and
 391 1D V_P^{tomo} models resampled in depth according to the V_S^{sw} layering (V_P^{rs}) (red solid line). V_S^{sw}
 392 models provide the best fit with the picked dispersion curves and the coherent maxima observed on
 393 dispersion images, supporting the validity of the final V_S^{sw} model averaged from all models fitting
 394 the error bars. For their part, dispersion curves computed from V_S^{tomo} models are generally not well
 395 fitting the observed propagation modes at low frequency, leading us to question the validity of the
 396 tomographic model in the deepest layers.

397 After cross-validating both V_S models, we computed V_P/V_S ratios along the line with: (i) V_S^{sw}
 398 and V_P^{rs} (Fig. 13a) and (ii) V_S^{tomo} and V_P^{tomo} (Fig. 13b). The V_S^{tomo}/V_P^{tomo} section shows smooth
 399 lateral variations, with low V_P/V_S (~ 1.5) in the western (from 0 m to 150 m) and central
 400 (from 200 m to 250 m) parts, separated by intermediate values ($\sim 2-2.5$). The eastern part is
 401 characterized by higher values (around 2.5 and up to 3.5). At first sight, the V_S^{sw}/V_P^{rs} section might
 402 look different, especially in the beginning of the line (from 0 m to 275 m), where anomalous high
 403 V_P/V_S values are observed around 5-m to 10-m depth. At these depths, the V_S^{sw} model presents
 404 mostly constant velocities, whereas the V_P^{rs} model is characterized by linearly increasing velocities.
 405 This incompatibility can thus explain the V_P/V_S discrepancies observed in this layer. With this in
 406 mind, we were yet able to delineate different V_P/V_S areas, which correspond well to those observed
 407 on the V_S^{tomo}/V_P^{tomo} section. These different areas and the observed V_P/V_S values are also consistent
 408 with the main structures delineated in the ERT results (Fig. 2b), whereas it was not clear on V_P or
 409 V_S only.

410 Moreover, 1D V_P/V_S ratios extracted at $X_{mid} = 50$ m (Fig. 12b) and $X_{mid} = 402$ m (Fig. 12e)
 411 from V_P^{rs}/V_S^{sw} (red solid line) and V_P^{tomo}/V_S^{tomo} (green solid line) show similar trends. Stronger
 412 contrasts and higher ratio values are yet observed on surface-wave analysis results. V_P/V_S values

413 observed at $X_{mid} = 50$ m are overall low (below 2.75) with both methods. The water table level
 414 extrapolated from the nearest representative piezometric well implanted in the granite (around
 415 100 m west from $X_{mid} = 50$ m) is not consistent with any of the contrasts observed on the V_P/V_S
 416 models. Indeed, the estimated water table level (~ 20 m) is close to the maximum HSD where
 417 V_S^{sw} becomes poorly resolved. Furthermore, no strong V_P/V_S variations can be anticipated in
 418 such low-permeability and low-porosity materials (Takei 2002). At $X_{mid} = 402$ m, shallow high
 419 V_P/V_S ratio values (around 4) are consistent with a wet soil, whereas a strong contrast at 11.5-m
 420 depth remarkably matches the water table level interpolated from levels measured in MF2 and F36
 421 (black dashed line). This feature is confirmed on the pseudo-2D V_P/V_S section in the eastern part
 422 (Fig. 13a).

423 6. DISCUSSION AND CONCLUSIONS

424 In order to assess the applicability of combined P-wave refraction tomography and surface-wave
 425 dispersion analysis to estimate V_P/V_S ratio in near-surface applications, we performed seismic
 426 measurements on a well-known granite–micaschists contact at Plœmeur hydrological observatory
 427 (France). A simultaneous P- and surface-wave survey was achieved using a single acquisition set-up
 428 and was supplemented with an SH-wave acquisition along the same line in order to compare V_S
 429 results obtained from SH-wave refraction tomography and surface-wave profiling.

430 P- and SH-wave first arrivals observed along the line were used to perform travel-time tomography
 431 and retrieve V_P^{tomo} and V_S^{tomo} models. Evenly spaced dispersion data were extracted along the line
 432 from P-wave shot gathers using windowing and stacking techniques. Successive 1D Monte Carlo
 433 inversions of these dispersion data were achieved using fixed V_P values extracted from the V_P^{tomo}
 434 model and no lateral constraints between two adjacent 1D inversions. The resulting 1D V_S^{sw} models
 435 were then assembled to create a pseudo-2D V_S^{sw} section. We computed normalized residuals between
 436 observed and calculated phase velocities along the line to control the quality of the V_S^{sw} model.
 437 This model appears to be correctly matching the V_S^{tomo} model obtained with SH-wave refraction
 438 tomography. The V_S^{sw} model is however characterized by strong velocity uncertainties in the deepest
 439 layers. The recomputation of theoretical dispersion curves along the line also provided results that
 440 are consistent with the measured dispersion images and proved to be a reliable tool for validating
 441 V_S models obtained from SH-wave refraction tomography and surface-wave profiling. Finally, we

442 were able to compute V_P/V_S sections from both V_S^{sw} and V_S^{tomo} . The two sections present similar
443 features, but the section obtained from V_S^{sw} shows a higher lateral resolution, which is consistent
444 with the features observed on ERT, thus validating our approach for retrieving V_P/V_S ratio from
445 combined P-wave tomography and surface-wave profiling. Furthermore, the V_P/V_S ratios obtained
446 in the clay and micaschists area show a strong contrast consistent with the observed water table
447 level.

448 An incompatibility, however, remains between V_P^{rs} and V_S^{sw} , which can lead to anomalous V_P/V_S
449 values. Indeed, travel-time tomography provides a smooth 2D reconstruction of the medium along
450 ray paths propagating from sources to sensors in a narrow frequency band. The investigation depth
451 is strongly related to the length of the acquisition profile and the maximum offset between sources
452 and receivers. Furthermore, the medium is described as a function of the ray coverage, which is
453 strongly related to the spacing between sensors and usually increases in high-velocity zones. For
454 their part, surface-wave methods allow reconstructing pseudo-2D V_S sections by juxtaposing single
455 1D models obtained along the line. In this case, the spectral resolution and the investigation depth
456 are a function of the frequency and increase with the length of the recording array, whereas the
457 lateral resolution is inversely correlated with the individual array length. In order to retrieve V_P and
458 V_S models more suited for V_P/V_S computation, a joint inversion approach combining Rayleigh and
459 P-guided waves dispersion data along with P-wave refraction travel times could be developed, in
460 the continuation of [Maraschini *et al.* \(2010\)](#), [Piatti *et al.* \(2013\)](#) and [Boiero, Wiarda, and Vermeer
461 \(2013b\)](#) works.

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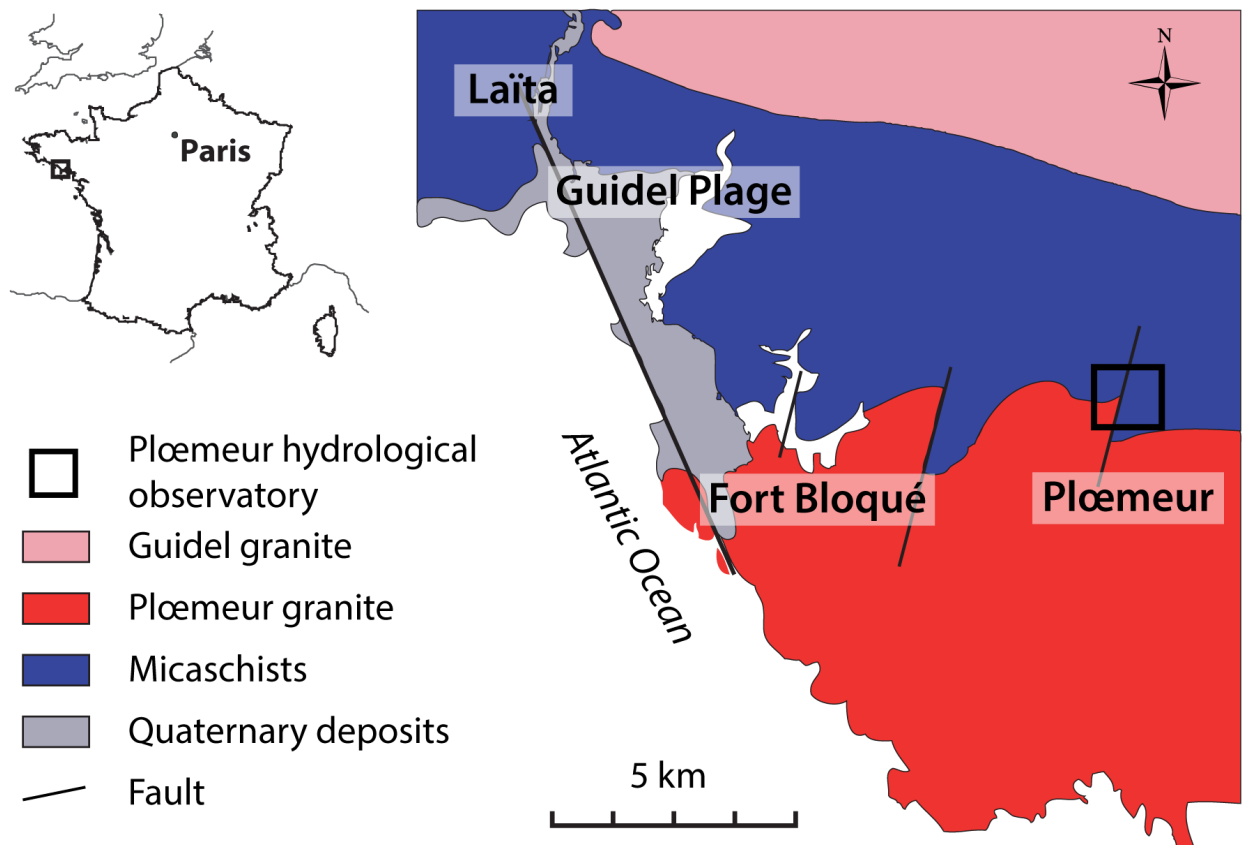


Figure 1: Geological context of the Plœmeur hydrological observatory and location of the experimentation (modified from Ruelleu *et al.* (2010) and Jiménez-Martínez *et al.* (2013)).

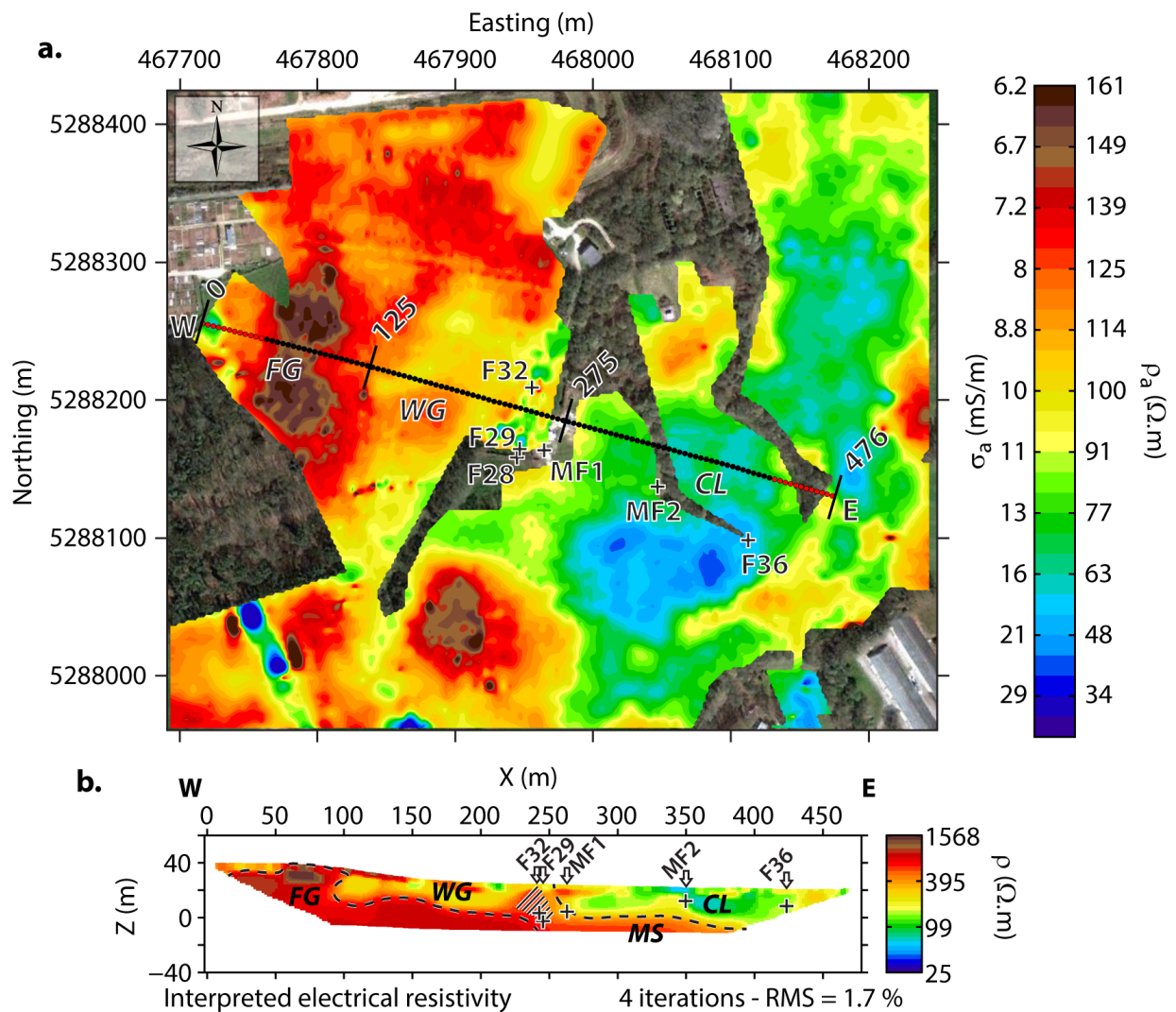


Figure 2: (a) Apparent EC values (σ_a) obtained from slingram EC mapping and location of the acquisition line (WE) for ERT and seismic measurements. Black dots between 30 m and 446 m represent the extent of the surface-wave analysis, whereas red dots represent the full extent of the survey. Locations of monitoring wells (F* and MF*) are represented with white crosses. (b) Electrical resistivity values (ρ) interpreted from ERT carried out along the WE line. Four main structures are delineated: fresh granite (FG), weathered granite (WG), clays (CL), and micaschists (MS). Markers at 125 m and 275 m in (a) delineate the three main shallow structures. The hashed area in (b) corresponds to a possible evidence of the contact zone. Positions of the nearest monitoring wells are projected along the WE line in (b) and represented with white arrows, pointing downwards to the corresponding piezometric head level (black crosses).

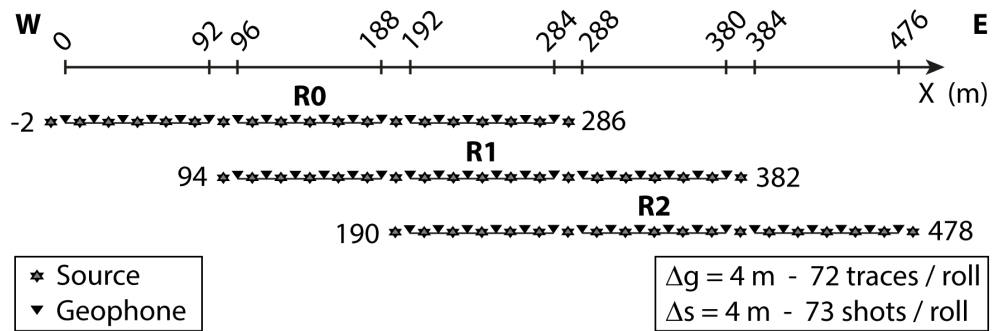


Figure 3: Seismic acquisition set-up used for combined P-, surface-, and SH-wave surveys. P- and surface-wave data were obtained using 72 14-Hz vertical geophones, whereas SH waves were recorded with 72 14-Hz horizontal geophones. Interval between two geophones (Δg) and move-up between shots (Δs) were both 4 m. We used one base roll (R0) and two roll-alongs (R1 and R2) of 72 traces each, with an overlap of 48 traces between two successive rolls, to obtain a 476-m-long profile. The origin of the x -axis is identical to the one used for ERT (Fig. 2b).

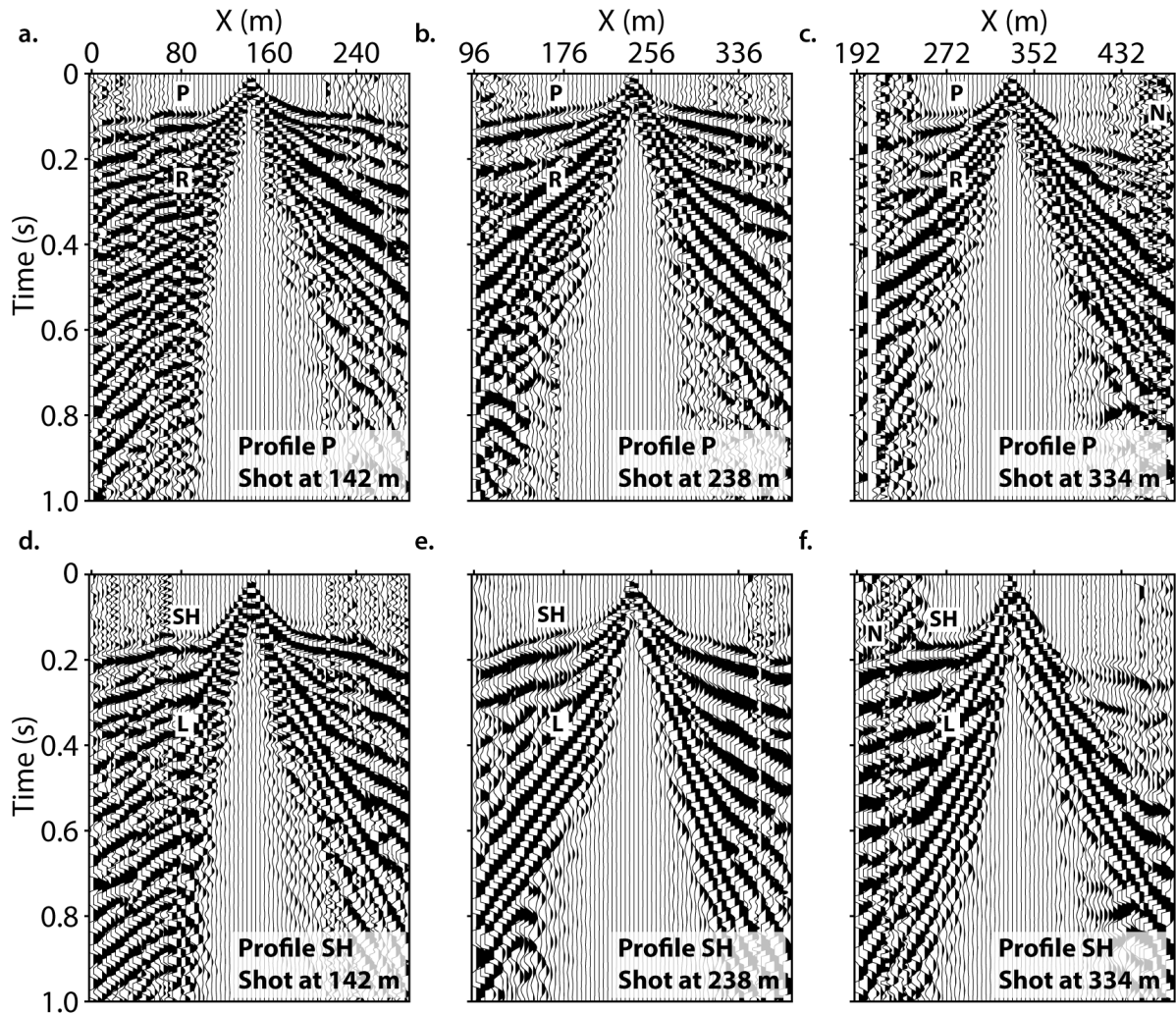


Figure 4: Seismograms for three shots recorded with vertical component geophones (profile P) and located at the centres of the base roll (a, $x = 142$ m), the first roll-along (b, $x = 238$ m), and the second roll-along (c, $x = 334$ m). Seismograms for three shots recorded with horizontal component geophones (profile SH) and located at the centres of the base roll (d, $x = 142$ m), the first roll-along (e, $x = 238$ m), and the second roll-along (f, $x = 334$ m). Data were only clipped over the 70th percentile in amplitude. P-wave (P) and Rayleigh-wave (R) are observed on seismograms recorded with vertical geophones. SH-wave (SH) and Love-wave (L) are visible on seismograms recorded with horizontal geophones. Data are affected with significant noise level (N) at far offset.

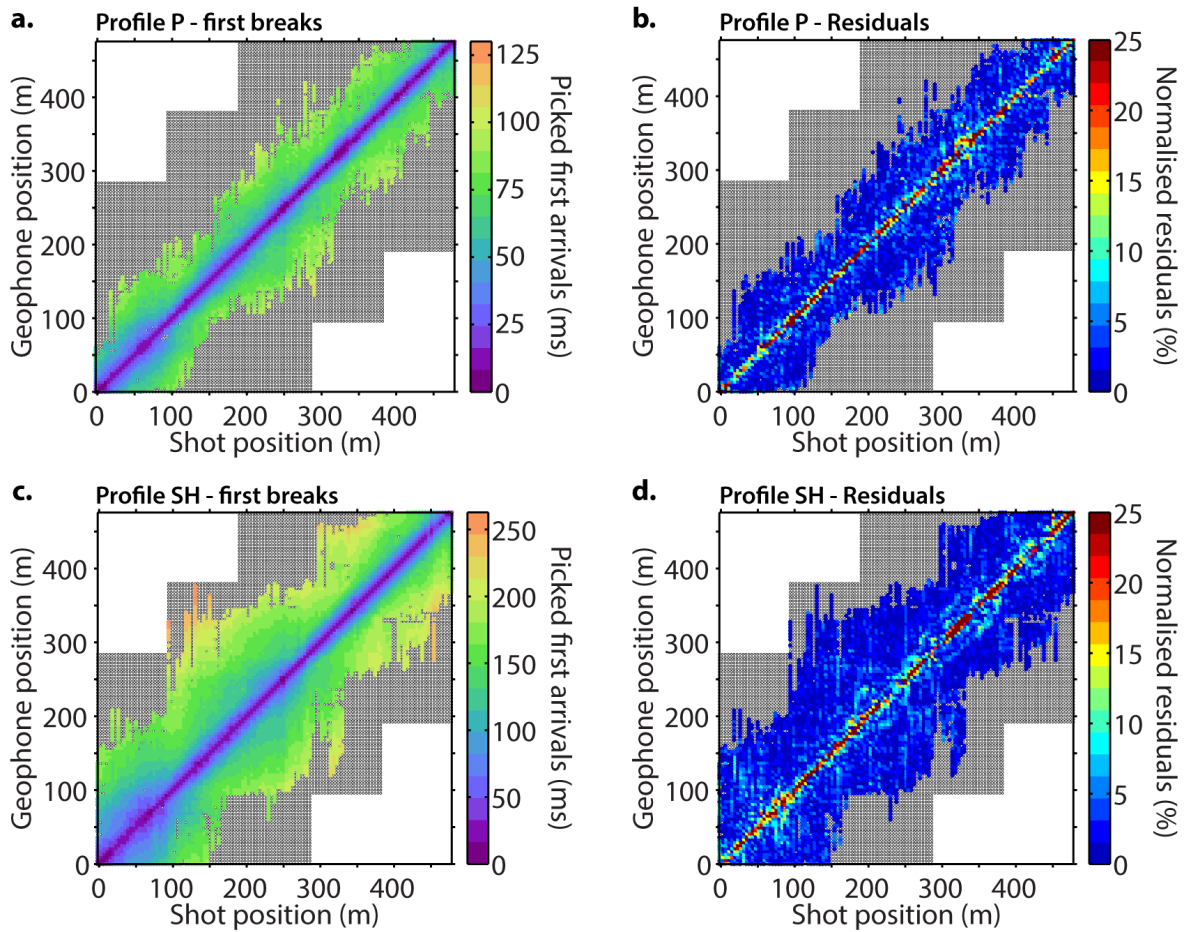


Figure 5: P-wave (a) and SH-wave (c) picked first arrival times and corresponding normalized residuals computed after P-wave (b) and SH-wave (d) tomography inversions, represented as a function of source and receiver distance. Black dots correspond to existing traces, whereas coloured dots represent picked traces. The diagonal trace (equal source and receiver location) indicates zero-offset travel times.

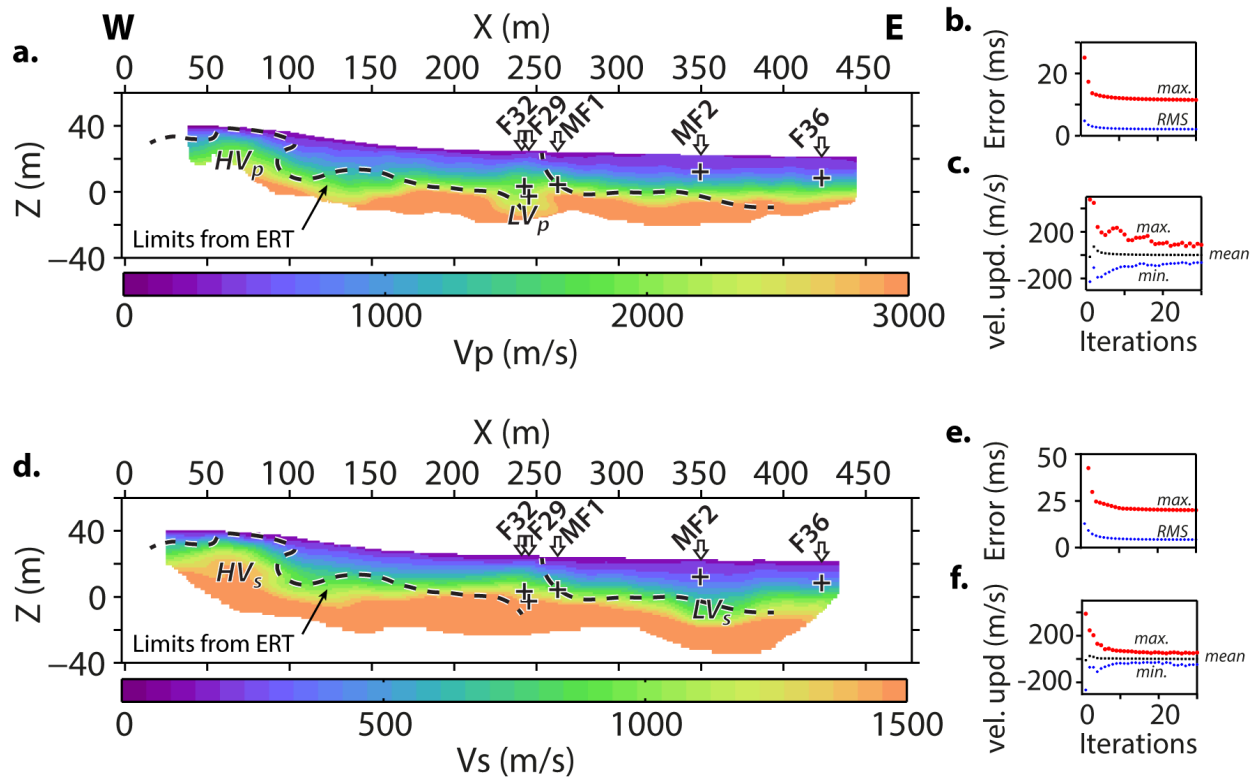


Figure 6: Final V_P (a) and V_S (d) models obtained after a total of 30 iterations, overlaid with limits interpreted from ERT results. The velocity models are clipped below a ray coverage of 100 rays per grid cell. The main features observed on the models are: (i) higher V_P and V_S in the western part (HV_P and HV_S , respectively); (ii) lower V_P in the centre (LV_P); and (iii) lower V_S in the eastern part of the profile (LV_S). Changes of maximum normalized residual (in red) and global RMS error (in blue) are represented along the inversion process for V_P (b) and V_S (e) models. Maximum, minimum, and mean velocity updates (in red, blue, and black, respectively) are represented at each inversion iteration for V_P (c) and V_S (f) models. Positions of the nearest monitoring wells are projected along the WE line as in Fig. 2b.

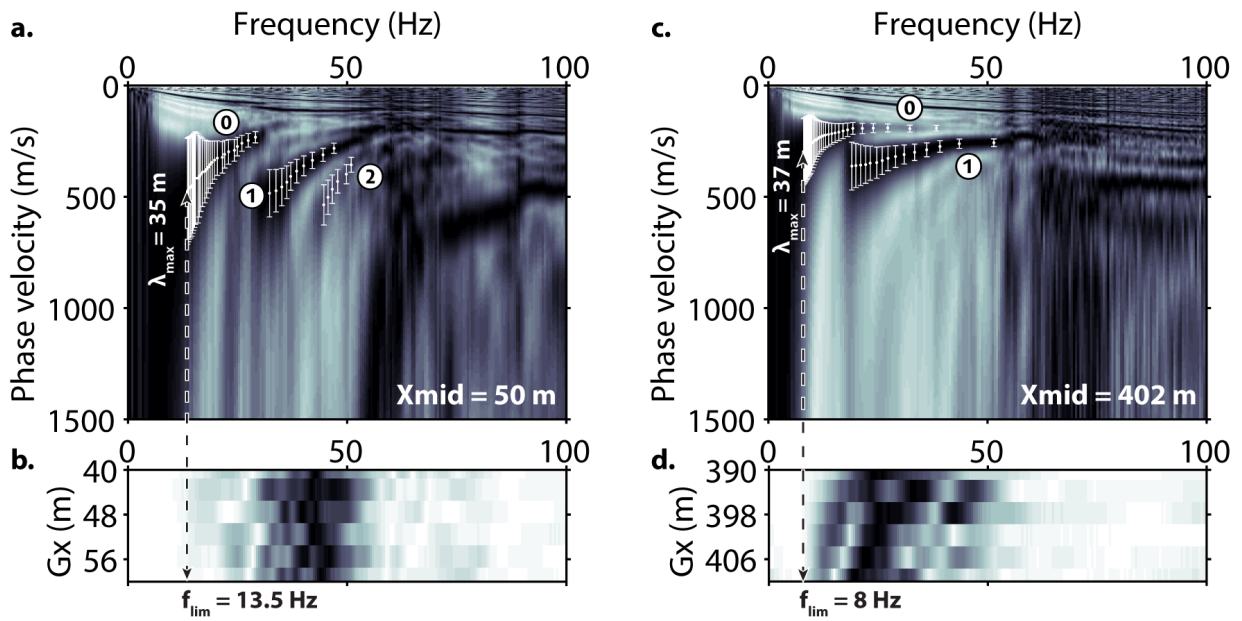


Figure 7: Dispersion images extracted on both sides of the seismic profile (a, $X_{mid} = 50$ m, and c, $X_{mid} = 402$ m) from six direct and six reverse shots with a six-trace (20-m-wide) window. X_{mid} is the position of the spread mid-point along the line. Spectral amplitude of the far offset direct shot is represented for both examples (b, $X_{mid} = 50$ m, and d, $X_{mid} = 402$ m). Picked dispersion curves are represented for the fundamental (0), first (1), and second (2) higher modes, with error bars defined according to the workflow described in O'Neill (2003). Dispersion curves were limited down to the frequency (f_{lim}) where the spectral amplitude of the seismogram became too low. The corresponding maximum wavelength (λ_{max}) was also extracted to indicate typical investigation depth criterion.

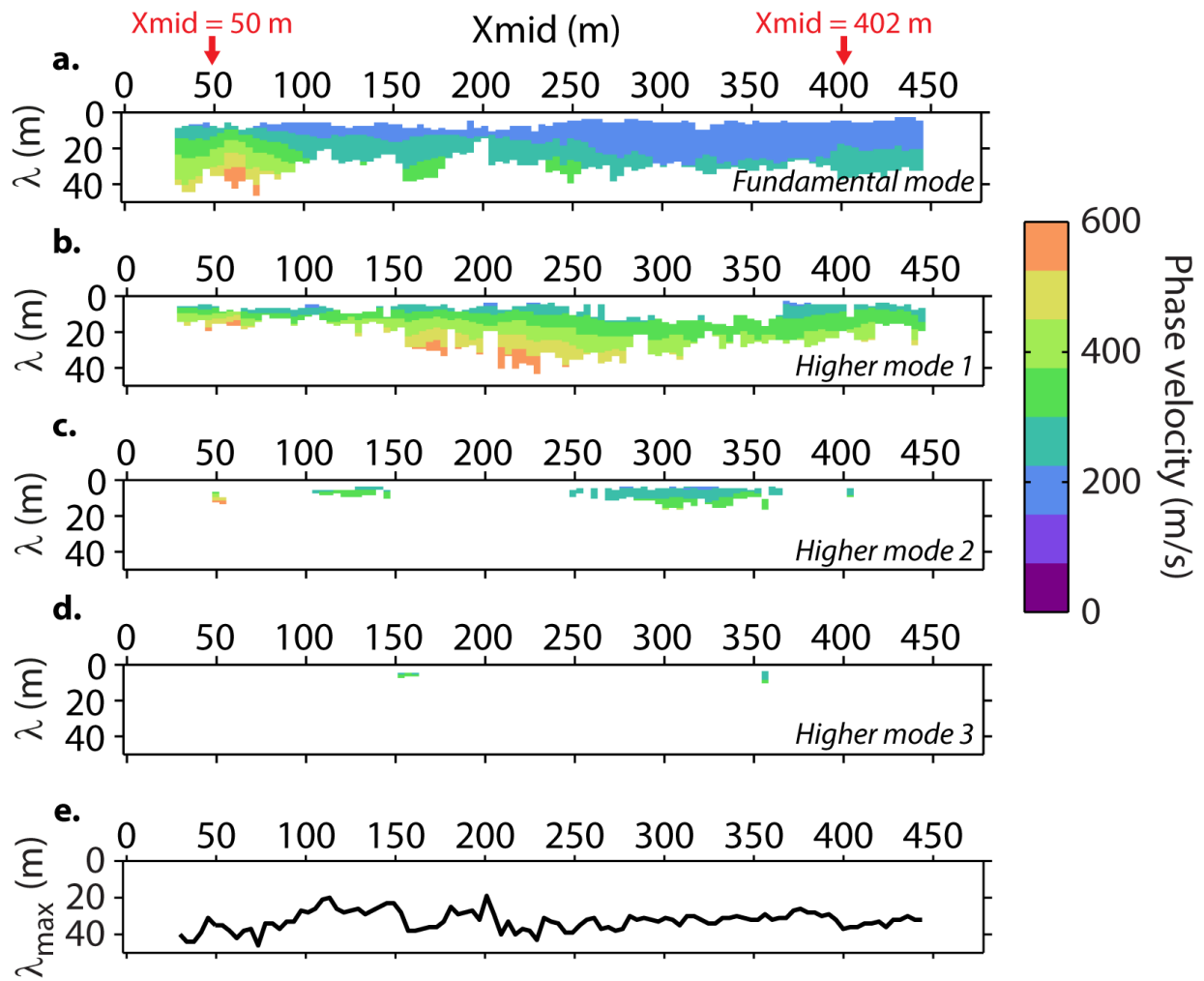


Figure 8: Pseudo-section of Rayleigh-wave phase velocity dispersion curves picked for the fundamental (a), first (b), second (c), and third (d) higher modes along the line after dispersion stacking, represented as a function of the wavelength (λ) and the spread mid-point (X_{mid}). (e) Maximum wavelength (λ_{max}) observed at each X_{mid} .

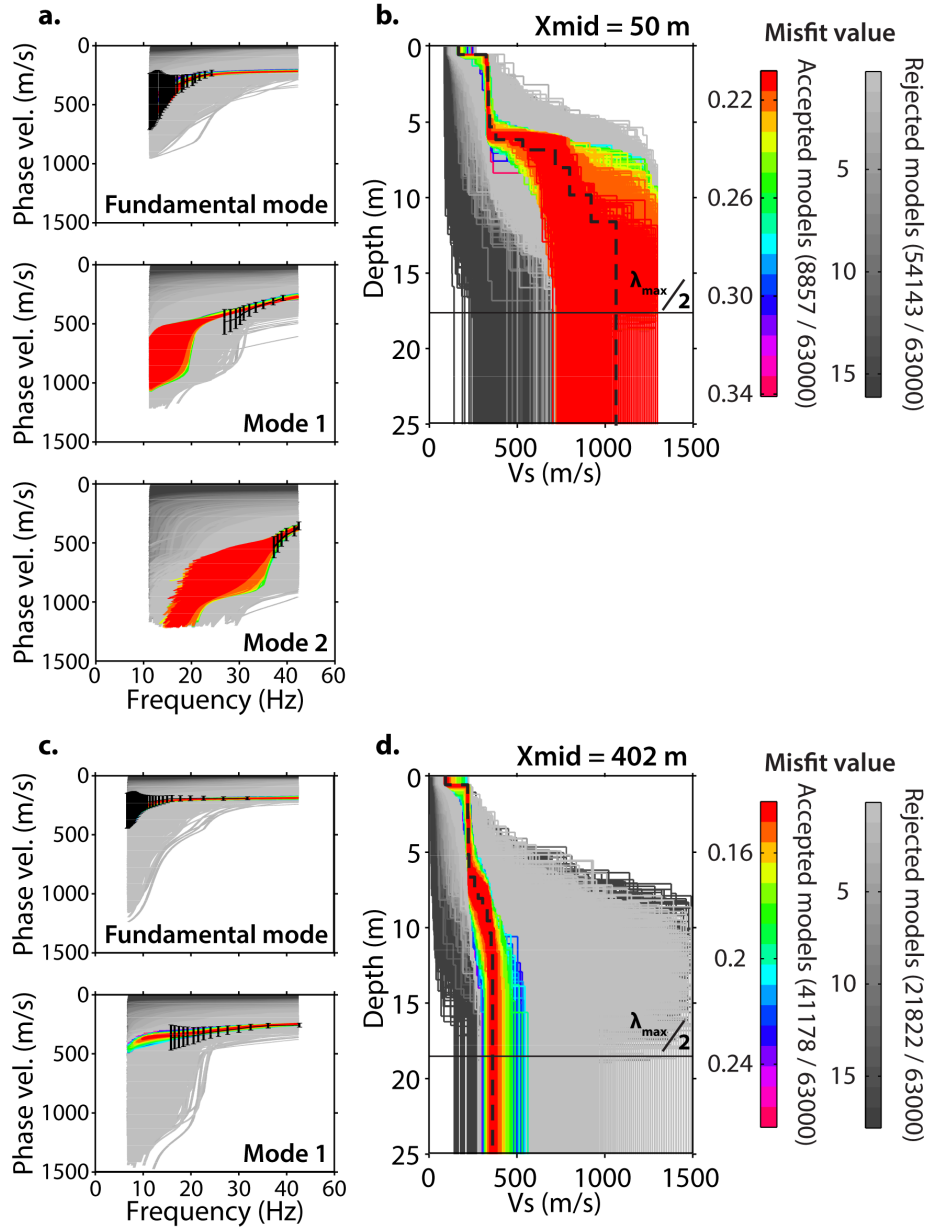


Figure 9: One-dimensional inversion of dispersion data (black error bars) extracted from the stacked dispersion image at $X_{mid} = 50$ m (a) and $X_{mid} = 402$ m (c) using the NA as implemented by [Wathelet et al. \(2004\)](#). Resulting models are represented for $X_{mid} = 50$ m (b) and $X_{mid} = 402$ m (d). Rejected models (i.e., at least one sample of the theoretical dispersion curves calculated from the model does not fit within the error bars) are represented according to their misfit with a greyscale, whereas accepted models (i.e., all samples of the theoretical dispersion curves calculated from the model fit within the error bars) are represented with a colour scale. Average parameters of all accepted models were used to build an average velocity structure associated with the centre of the extraction window (black dashed lines in b and d). The black solid line in (b) and (d) corresponds to $\lambda_{max}/2$ and indicates a typical investigation depth criterion.

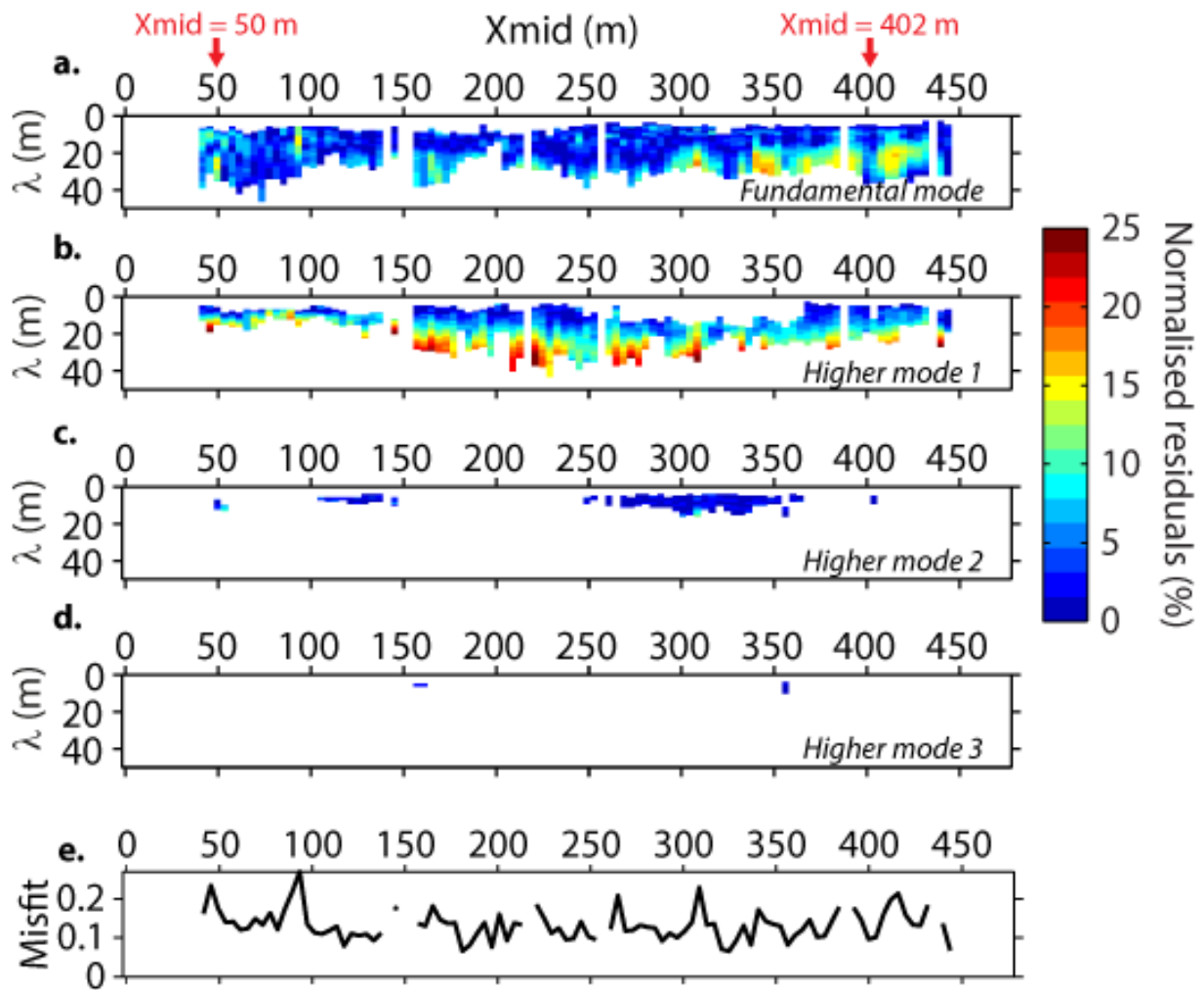


Figure 10: Pseudo-section of Rayleigh-wave phase velocity normalized residuals for the fundamental (a), first (b), second (c), and third (d) higher modes along the line after dispersion stacking. λ is the wavelength. (e) Misfit values calculated with equation (1) for each 1D inversion along the line.

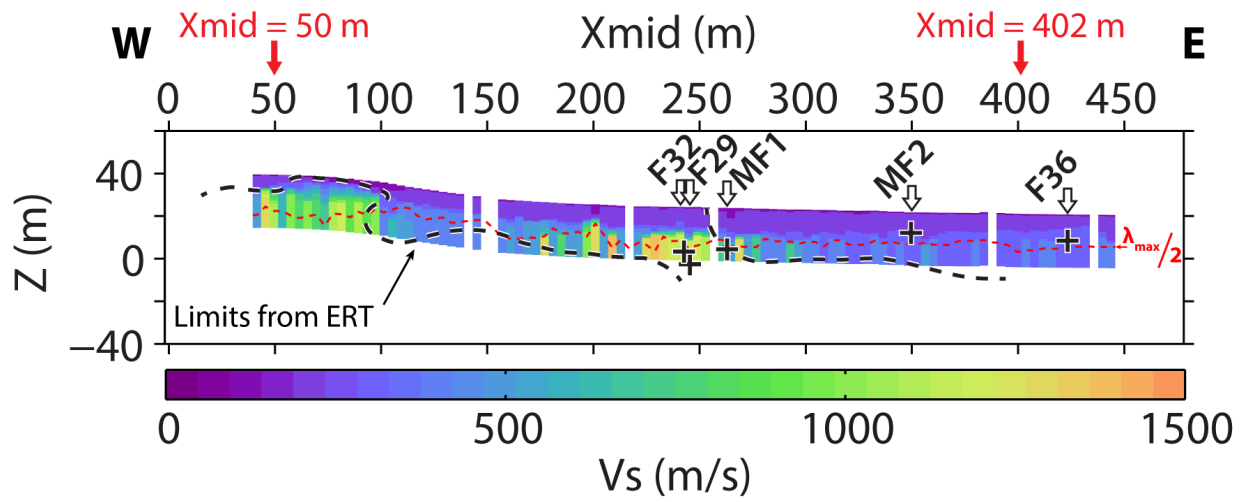


Figure 11: Pseudo-2D V_S section constructed with all 1D V_S models obtained from surface-wave profiling, overlaid with limits interpreted from ERT results. $\lambda_{max}/2$, represented with a black dashed line, corresponds to a typical investigation depth criterion. Positions of the nearest monitoring wells are projected along the WE line as in Fig. 2b.

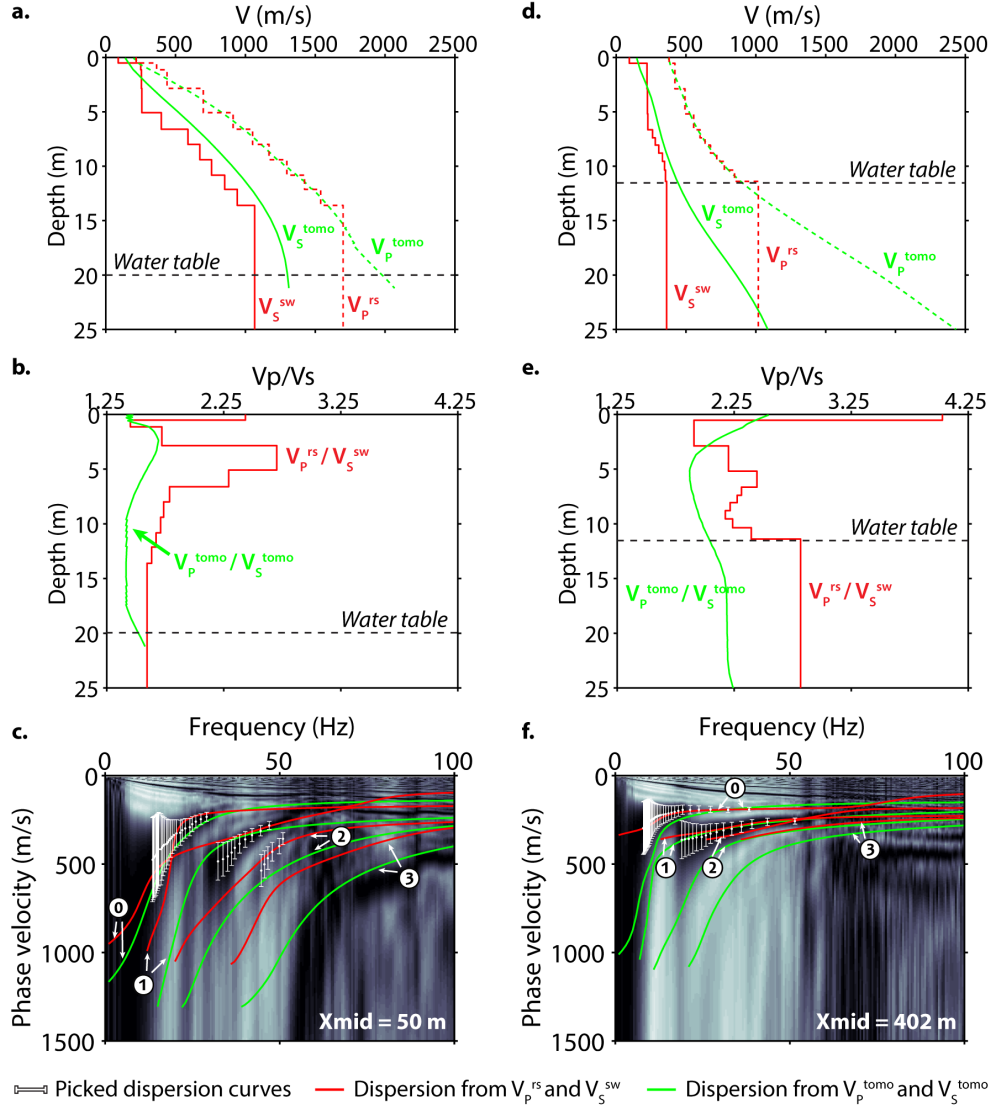


Figure 12: One-dimensional V_S models obtained from SH-wave refraction interpretation (V_S^{tomo} , green solid line) and surface-wave dispersion inversion (V_S^{sw} , red solid line) at $X_{mid} = 50$ m (a) and $X_{mid} = 402$ m (d). One-dimensional V_P models obtained from P-wave refraction tomography (V_P^{tomo} , green dashed line) and resampled to follow the layering of V_S^{sw} (V_P^{rs} , red dashed line) are also represented in (a) and (d). One-dimensional V_P/V_S models constructed from V_S^{sw} and V_P^{rs} (V_P^{rs}/V_S^{sw} , red solid line) and from V_P^{tomo} and V_S^{tomo} (V_P^{tomo}/V_S^{tomo} , green solid line) at $X_{mid} = 50$ m (b) and $X_{mid} = 402$ m (e). The water table level is represented with a black dashed line. In (a) and (b), the water table level is extrapolated from the nearest representative well implanted in the granite (around 100 m west from $X_{mid} = 50$ m). In (d) and (e), the water table level is interpolated from levels measured in wells MF2 and F36 (Fig. 2). Stacked dispersion image obtained at $X_{mid} = 50$ m (c) and $X_{mid} = 402$ m (f). The fundamental (0), the first (1), second (2), and third (3) higher modes' dispersion curves calculated from V_S^{sw} and V_P^{rs} (in red) and from V_S^{tomo} and V_P^{tomo} (in green) are superimposed on both images.

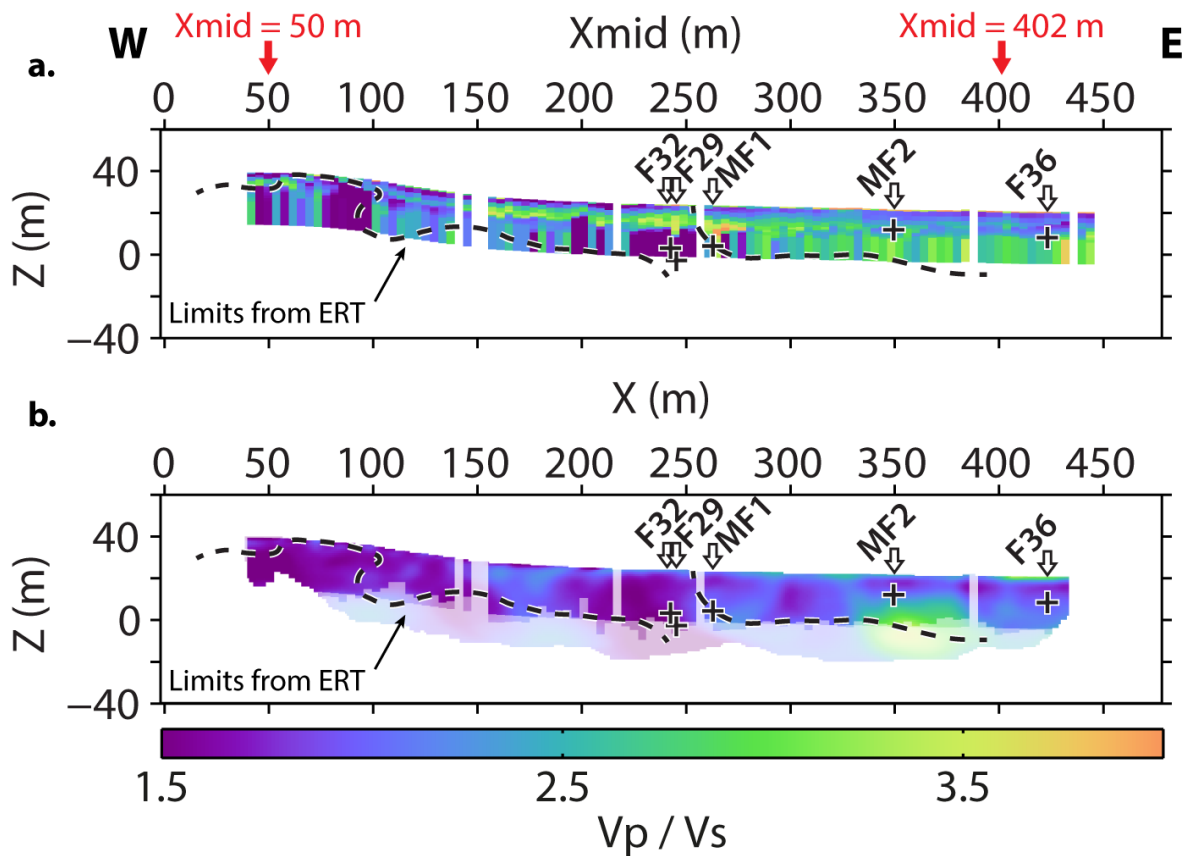


Figure 13: (a) Pseudo-2D V_P/V_S section constructed with V_S^{sw} obtained from surface-wave profiling and V_P^{rs} obtained from P-wave tomography and resampled in depth according to the V_S layering. (b) V_P/V_S section computed from V_P^{tomo} and V_S^{tomo} models obtained from P- and SH-wave tomography. The shaded area in (b) corresponds to the extent of the V_P/V_S pseudo-2D section in (a). Both sections are overlaid with limits interpreted from ERT results. Positions of the nearest monitoring wells are projected along the WE line as in Fig. 2b.