



1 Depth-dependent seismic anisotropy in the oceanic crust of the Nazca Plate **2 offshore northern Chile**

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6 Abstract

7 Seismic anisotropy preserves information about the tectonic, structural, and hydrothermal evolution of
8 oceanic lithosphere, yet direct observations from mature oceanic crust remain sparse. Here, we present
9 constraints on azimuthal P-wave anisotropy from an active source seismic experiment conducted on a
10 ~45 Ma old segment of the Nazca Plate at ~25.5°S before the plate is affected by outer-rise bending
11 prior to its subduction. Traveltime residuals show a coherent twofold azimuthal pattern for rays with
12 maximum penetration depths within the upper ~2 km of the crust, with faster propagation
13 approximately E-W and slower propagation approximately N-S. Apparent anisotropy inferred from the
14 analysis reaches ~4.5 - 5.5 % for the shallowest ray group down to 1 km below seafloor, decreasing to
15 3 - 4 % for rays penetrating to ~1 - 2 km depth. Below ~2 km depth, in the lower crust, no coherent
16 anisotropic signature is resolved. The fast symmetry axis strikes obliquely to the paleo-spreading
17 fabric and is parallel to E-W oriented horst and graben structures related to N-S extension. We
18 therefore interpret the observed azimuthal velocity variation as being primarily associated with the
19 younger fault and fracture architecture. The weakening of the coherent anisotropic signal with
20 increasing ray penetration depth is consistent with decreasing fracture compliance and /or density with
21 depth, potentially related to increasing confining pressure and progressive mineral sealing. To place
22 these observations in a broader geodynamic context, a comparison with previous oceanic-crust studies
23 indicates that sampling depth and tectonic fabric strongly influence reported anisotropy, whereas the
24 available data are insufficient to isolate a robust dependence on plate age or spreading rate.

25 Short summary

26 In this study, we used active-source seismic data to investigate crustal anisotropy in a 45 Ma old
27 segment of the Nazca Plate. We identified a strong twofold anisotropic pattern in the upper crust, with
28 fast wave propagation in E-W direction. The anisotropy decreases with depth and is best explained by
29 fractures associated with horst and graben structures. Our results highlight how tectonic deformation
30 impacts seismic properties and anisotropic structure of oceanic crust.

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33 1 Introduction

34 The oceanic lithosphere records seafloor spreading, tectonic deformation, magmatic accretion,
35 and hydrothermal alteration (e.g., Parsons and Sclater, 1977; White et al., 1992; Cannat, 1993;
36 Grevemeyer and Weigel, 1996). Seismic anisotropy is particularly sensitive to these processes as it
37 reflects the orientation and abundance of cracks and faults, the magmatic fabric, and the mineral
38 alignment (e.g., Crampin, 1981; Mainprice, 2010). The general structure and anisotropic behavior of
39 the oceanic lithosphere is known since the 1970s and 1980s and has been investigated by several
40 authors (e.g., Furumoto et al., 1971; Stephen, 1981, 1985; White and Whitmarsh, 1984; Shearer and
41 Orcutt, 1985, 1986). In the upper mantle, anisotropy is commonly related to the lattice-preferred
42 orientation of mantle minerals, such as olivine and pyroxene, which is induced by mantle flow and
43 deformation. Contrarily, in the oceanic crust, azimuthal anisotropy is often attributed to aligned ridge-
44 parallel or subparallel cracks and fissures that formed during crustal accretion and extension at
45 spreading centers (e.g., Stephen, 1985; Almendros et al., 2000; Becker et al., 2014; Russell et al.,
46 2019). Despite this general understanding, direct observations from shallow oceanic crust remain
47 comparatively rare and are concentrated on young oceanic lithosphere near Mid-Ocean-Ridges (e.g.,
48 White and Whitmarsh, 1984; Dunn and Toomey, 2001), whereas other studies predominantly target
49 the upper oceanic mantle (e.g., Becker et al., 2014; Lin et al., 2016; Russell et al., 2019) or the
50 subducting slab beneath the marine and continental forearc using land-based observations (e.g.,
51 MacDougall et al., 2012; Agrawal et al., 2020; Condori et al., 2021). Consequently, the extent to
52 which anisotropic fabrics, shaped at spreading centers, persist, are sealed, or are overprinted by
53 younger intraplate deformation remains poorly constrained in mature oceanic crust. This question is
54 particularly relevant for incoming plates because fault and fracture fabrics can influence permeability,
55 fluid distribution, and stress transfer prior to subduction (e.g., Crampin, 1994; Crampin and Peacock,
56 2005; Long and Silver, 2008; Faccenda et al., 2009). For the Nazca Plate offshore South America,
57 available constraints are sparse and heterogeneous: Stephen (1981, 1985) reported up to 20 % crustal
58 anisotropy within the uppermost 500 m at a borehole offshore Colombia, whereas Contreras-Reyes et
59 al. (2008) used crossing P-wave velocity profiles to detect less than 2 % of seismic anisotropy in the
60 upper mantle at the South Chilean outer rise at ~43°S. The central segment of the Nazca Plate
61 therefore provides an opportunity to test how anisotropy in mature incoming crust relates to inherited
62 spreading fabric and younger fault systems.

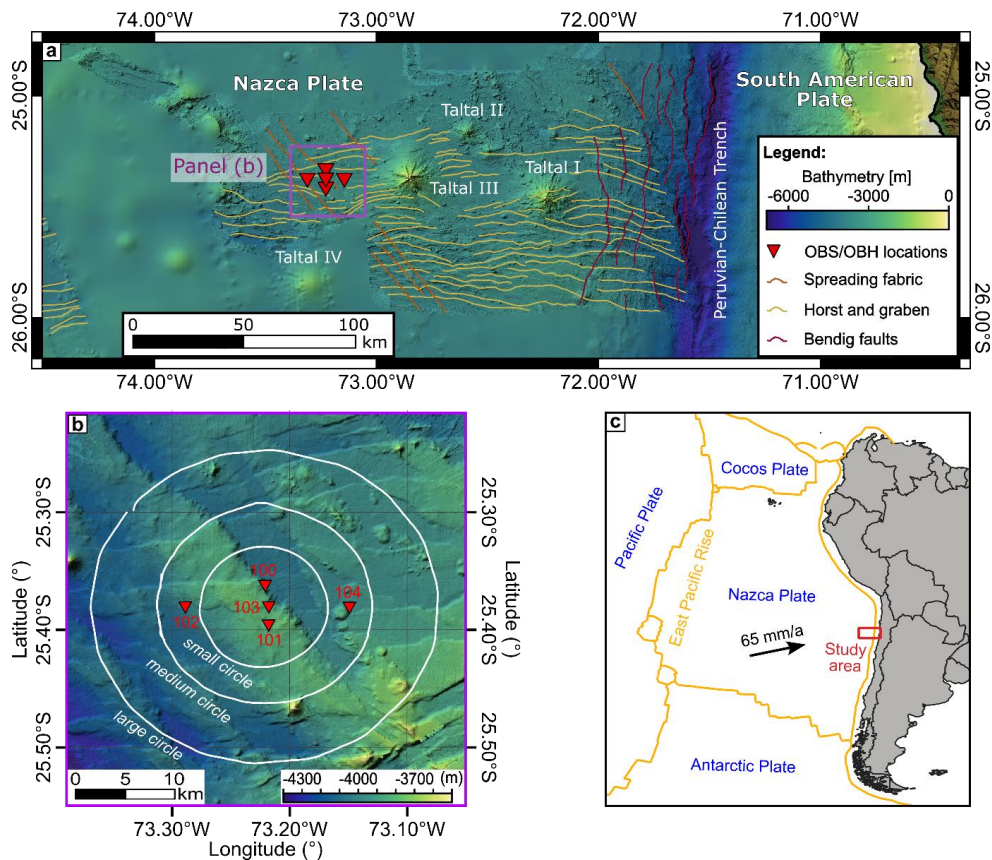
63 In this study, we analyze active seismic refraction data recorded on Ocean-Bottom-
64 Seismometers (OBS)/Ocean-Bottom-Hydrophones (OBH) to investigate crustal anisotropy in a ~45
65 Ma old segment of the Nazca Plate at ~25.5°S that is still largely unaffected by plate bending towards
66 the subduction zone (Fig. 1a). The P- and S-wave velocity structure and the crustal geometry of the
67 region were recently determined by Warwel et al. (2025b). We use data from the same experiment
68 (*R/V SONNE SO297*; Lange et al., 2023), acquired along three circular source tracks surrounding five



69 OBS/OBH stations (Fig. 1b). This geometry provides dense azimuthal sampling and allows testing for
70 systematic variations in Pg traveltimes residuals. We investigate (1) whether a coherent azimuthal P-
71 wave anisotropic signal is resolved, (2) how the signal changes for rays with different maximum
72 penetration depth, and (3) whether the fast direction is more closely related to the inherited paleo-
73 spreading fabric or to younger fault systems mapped by Warwel et al. (2025b). We then place the
74 results in the context of previous observations of seismic anisotropy in oceanic crust.

75 **2 Tectonic Setting of the Nazca Plate**

76 The oceanic Nazca Plate, located west of the South American continental margin, exhibits a
77 complex geological history characterized by multiple phases of plate boundary reorganization,
78 microplate formation and destruction, as well as rifting events (e.g., Herron, 1972; Handschumacher,
79 1976; Hey et al., 1985; Searle, 1989; Hey et al., 1995; Tebbens and Cande, 1997). Emerging from the
80 breakup of the ancient Farallon Plate at approximately 23 Ma ago (Barckhausen et al., 2001), the
81 Nazca Plate is bounded by the Cocos Plate to the North, the Antarctic Plate to the South, and the South
82 American Plate to the East (Fig. 1c). Its spreading center, located along the East Pacific Rise to the
83 West, ranks among the fastest on Earth, with full spreading rates exceeding 12 cm/a (DeMets et al.,
84 2010). The Nazca Plate subducts along the Peruvian-Chilean trench and reaches a maximum age of
85 ~50 Ma at ~20°S (Müller et al., 2008). The current convergence rate is ~66 mm/a at an angle of
86 N79.9°E, however, a progressive slowdown in the eastward plate motion of the Nazca Plate has been
87 documented over the last 3.16 Ma (DeMets et al., 2010; Jarrin et al., 2023).



88

89 **Figure 1:** (a) Bathymetric map of the study area. High-resolution multibeam bathymetry from SO297
90 (Warwel et al., 2025c), gaps filled with bathymetry from GEBCO (GEBCO Compilation Group,
91 2023). Lines represent faults and lineaments identified in the bathymetry by Warwel et al. (2025b). (b)
92 Close-up showing the anisotropy experiment indicating the OBS/OBH stations with three active
93 seismic acquisition circles. (c) Plate tectonic setting in the eastern Pacific region.

94 The Nazca Plate hosts significant structural heterogeneities, including seamount chains and
95 ridges with different orientations and ages, as well as extensive fault structures oriented either parallel
96 or perpendicular to the trench (Bello-González et al., 2018; Warwel et al., 2025b). Our study area at
97 ~25.40°S, -73.25°W is characterized by seamounts of the Taltal Ridge, E-W oriented horst and graben
98 structures and NW-SE oriented lineaments. The horst and graben structures are suggested to be related
99 to N-S extension within the last 10 Ma, while the NW-SE lineaments are interpreted to be related to
100 the paleo-spreading fabric (Warwel et al., 2025b) (Fig. 1a). The closest major seamount, Taltal III, is
101 located ~40 km to the East of the study area, however, smaller cones and seamounts are also visible
102 within the study area (Fig. 1b). The Nazca Plate exhibits a ~7 km thick oceanic crust with no
103 significant sediment cover. The velocity structure shows P-wave velocities in the upper ~2 km thick



104 crust between ~ 3 km/s and ~ 6 km/s and a high velocity gradient of > 1 s $^{-1}$. The lower crustal velocities
105 between ~ 6 km/s and ~ 7.2 km/s are paired with a smaller velocity gradient of $\sim 0.15 - 0.25$ s $^{-1}$, typical
106 for magmatically accreted oceanic crust (Grevenmeyer et al., 2018). For the entire crust, the V_p/V_s ratio
107 generally ranges from ~ 1.80 to ~ 1.95 , with the greatest variability in the upper crust (Warwel et al.,
108 2025b).

109 3 Data and Methods

110 During *R/V SONNE* expedition SO297 (PISAGUA) offshore northern Chile (Lange et al.,
111 2023), active wide-angle seismic refraction data were acquired as part of a regional-scale experiment
112 to investigate the subduction geometry and dynamics offshore the central Atacama region (Warwel et
113 al., 2025a, 2025b). The data in this study were recorded on four Ocean-Bottom-Seismometers (OBS)
114 and one Ocean-Bottom-Hydrophone (OBH), placed in a cross-shaped layout (Fig. 1b), at a sampling
115 frequency of 250 Hz. Seismic signals were generated by an array of 6 G-guns with a total volume of
116 ~ 40 l that fired every 30 s (~ 60 m distance) along three circles surrounding the stations with diameters
117 of 3, 5 and 8 nm (~ 5.5 , ~ 9.3 and ~ 14.8 km), respectively (Fig. 1b). Pre-processing of the seismic data
118 included cutting the continuously recorded signals into record sections for each station and converting
119 them into standard SEG-Y format, clock drift time correction, 3D relocation of the OBS/OBH
120 positions on the seafloor and bandpass filtering with corner frequencies of 2, 4, 20 and 40 Hz.

121 Investigating anisotropy in the oceanic crust can be conducted by analyzing P- and/or
122 converted S-waves. As the seafloor morphology results in inconsistent generation of S-waves along
123 the three acquisition circles as well as amplitude variations due to scattering, we base our analysis on
124 the first-arrival times of refracted P-waves (P_g). In a fully isotropic medium without lateral variations,
125 the traveltimes recorded at a station are independent of the azimuth angle between the source and the
126 station (e.g., Sil and Sen, 2008; Zhou and Greenhalgh, 2008). In contrast, an observed azimuth
127 dependence of traveltimes may indicate anisotropy within the medium. For all stations, P_g -arrival
128 times were picked for all three acquisition circles using *PasteUp* (Fujie et al., 2008) as a graphical
129 interface for *Seismic Unix* (Stockwell Jr, 1999) (Supplement Fig. S1), and the azimuths between shots
130 and stations were calculated. Subsequently, forward modeling of the ray paths and traveltimes for each
131 shot in 3D was conducted using the *TOMO3D* code from Meléndez et al. (2013, 2015). Thereby, we
132 utilized a 1D velocity-depth profile from Warwel et al. (2025b) that was extrapolated into three
133 dimensions with a grid size of 100-100-100 m (x-y-z) as an isotropic initial velocity model (Fig. 2b).
134 The resulting traveltime residuals between the P_g -picks and the modelled rays were then sorted into
135 different depth intervals according to the maximum depth of the respective rays (Fig. 2). Within these
136 depth intervals, all residuals with the same azimuth were binned into 1° bins, resulting in an average
137 traveltime residual for each azimuth angle and each depth interval (Fig. 3). To analyze the variation of
138 the residuals for different shot-receiver azimuths, we fitted a cosine function of the form
139 $\Delta t(\theta) = A * \cos(2(\theta - \theta_f)) + c$ (Δt = traveltime residual, A = amplitude, θ = azimuth, θ_f = azimuth

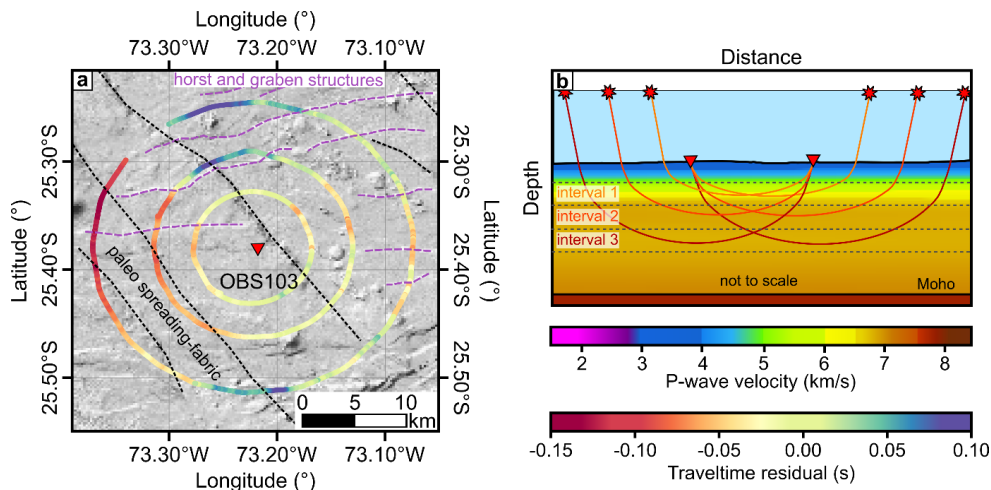


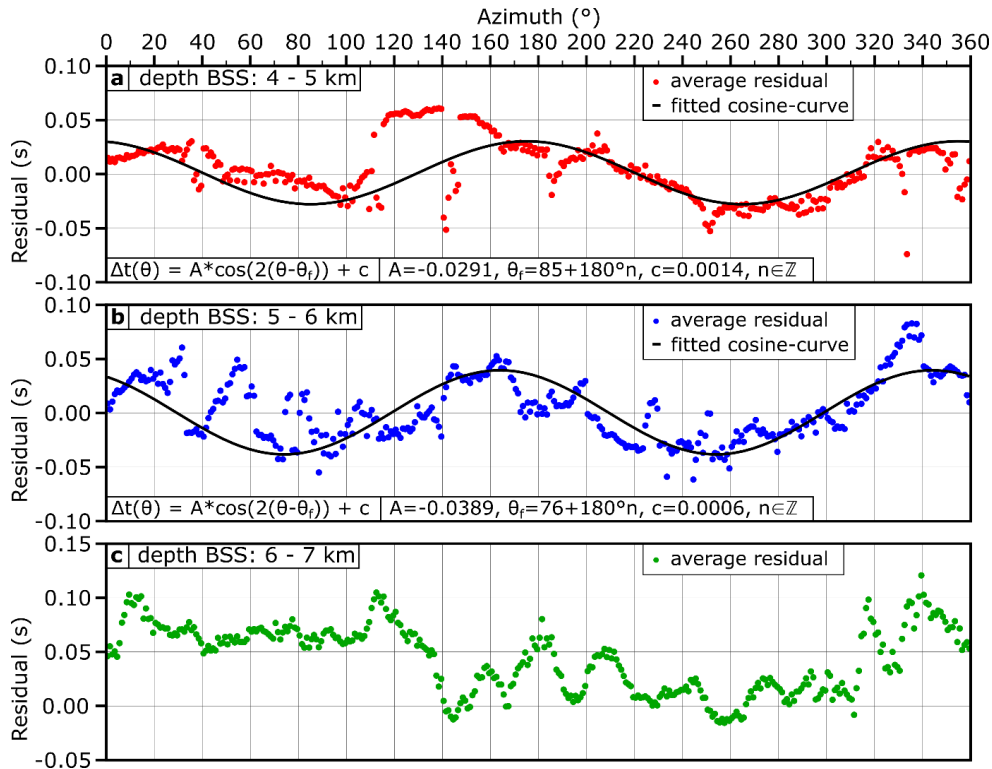
Figure 2: (a) Exemplary traveltime residuals for all acquisition circles at OBS103. SO297 bathymetry is displayed in gray. Black dotted lines represent lineaments related to the paleo-spreading fabric, dashed purple lines represent E-W oriented horst and graben structures (Warwel et al., 2025b). (b) Schematic illustration of ray paths and their sorting into different depth intervals based on their maximum penetration depth. The background colors illustrate the velocity model used for the ray tracing.

4 Results

The Pg traveltime residuals vary with source-receiver azimuth for rays with maximum penetration depths within the upper ~2 km of crust and show evidence of a systematic $\cos(2\theta)$ dependency (Fig. 3). Positive residuals indicate that traveltimes arrive later than predicted by the isotropic reference velocity model, whereas negative residuals indicate earlier arrivals. For the shallowest ray group with maximum depths of 4 - 5 km below sea surface (bss) (approximately 0 - 1 km below seafloor), residuals show two minima at ~85° and ~265° and two maxima at ~175° and ~355°, with a peak-to-peak difference of ~0.058 s (Fig. 3a). Rays with maximum penetration depth of 5 - 6 km bss (approximately 1-2 km below seafloor) show a similar twofold pattern, with minima at ~76° and ~256°, and maxima at ~166° and ~346°. The peak-to-peak difference is ~0.078 s (Fig. 3b). For both depth intervals, the uncertainties for the cosine coefficients are less than 20 % (Fig. S2). For the deepest investigated ray group, covering the transition from upper to lower crust with subsurface depths between 2 km and 3 km (depth bss: 6 - 7 km), the residuals remain variable but do not show a coherent twofold azimuthal pattern comparable to the shallower groups (Fig 3c). Traveltime residuals



163 > 0.05 s are observed at azimuths from ~0° to ~130° and ~320° to 360°, whereas between 130° and
 164 320° the residuals are mostly < 0.05 s (Fig. 3c).



166 **Figure 3:** Average traveltime residuals for different depth intervals (a: 4 – 5 km bss, b: 5 – 6 km bss,
 167 c: 6 – 7 km bss) plotted against the shot azimuth. An azimuth of 0° points towards North. The black
 168 line represents cosine curves fitted to the data. Depth bss: depth below sea surface.

169 5 Discussion

170 5.1 Degree of anisotropy and its symmetry axes

171 The shallow and intermediate ray groups in the upper crust show the same 2 θ -pattern:
 172 negative residuals approximately E-W and positive residuals approximately N-S. Within the deepest
 173 ray group located at the transition to the lower crust, a 1 θ -pattern occurs, with small residuals
 174 dominating the W and residuals undulating above 0.05 s in the E. A clear periodicity cannot be
 175 observed (Fig. 3). Based on this pattern, we interpret the residual variations in the upper crust (down to
 176 about 2 km below the seafloor) as an indication of crustal anisotropy within the Nazca Plate. We
 177 interpret two main axes of symmetry, with the fast axis striking approximately E-W and the slow axis
 178 striking approximately N-S. The exact strike directions show minimal variation between the two depth
 179 intervals and can be averaged to approximately 81° and 261° for the fast symmetry axis, and



approximately 171° and 351° for the slow symmetry axis, with azimuths measured clockwise from north (0°).

Dedicated active seismic anisotropy experiments focusing on shallow marine settings are relatively rare and often only consist of data derived from crossing refraction profiles (e.g., Contreras-Reyes et al., 2008, 2024). Nevertheless, crustal anisotropy in oceanic lithosphere is commonly associated with the oceanic spreading fabric (e.g., Furumoto et al., 1971; White and Whitmarsh, 1984; Stephen, 1985). These and other authors have linked crustal anisotropy to vertically aligned cracks and microfractures formed during crustal accretion, leading to higher velocities parallel to the spreading axis and lower velocities in the spreading direction, where seismic rays are impeded and slowed by these fractures (see Table 1). Our data show that the fast symmetry axis does not align with the paleo-spreading lineaments but strikes ~50° obliquely to them and rather aligns with E-W oriented horst and graben structures that have been observed by Warwel et al. (2025b) (Figs. 1a, 2a). We assume that the same structural mechanisms result in lower seismic velocities in the direction of extension (N-S) and higher velocities parallel to the faults (E-W). Furthermore, we propose that the alignment of the fast symmetry axis along the E-W horst and graben structures, rather than along the paleo-spreading fabric, reflects the relative ages of the fault structures. The paleo-spreading lineaments are approximately 45 Ma old, and the associated fractures and fissures have most likely undergone extensive alteration and sealing, minimizing the effect of potential anisotropy associated with these structures. In contrast, the fractures of the horst and graben structures evolved within the last 10 Ma (Warwel et al., 2025b), leaving substantially less time for alteration and sealing processes to occur. This suggests that fractures associated with horst and graben structures have overprinted a potential impact of the spreading fabric and currently exert dominant control over crustal anisotropy and remain at least partially mechanically open.

Table 1: Summary of marine crustal anisotropy studies, their degree of observed anisotropy as well as their interpretation / alignment with respect to the spreading axis.

Layer	Experiment type	Anisotropy	Interpretation / Direction	Reference
lower crust	active seismic	~7 %	fast direction parallel to the ridge axis	Furumoto et al. (1971)
upper crust	active seismic	~5 %	fast direction aligned approx. parallel to cracks	White and Whitmarsh (1984)
upper crust	active seismic	up to ~20 %	fast direction parallel to the spreading ridge	Stephen (1985)
upper crust	active seismic	~7 %	fast direction parallel to cracks and sheeted dykes	Shearer and Orcutt (1985)
upper crust	active and passive seismic	-	fast direction parallel to Juan de Fuca Ridge	Almendros et al. (2000)
upper crust	active seismic	up to 4 %	fast direction aligned ~6° to the spreading axis	Dunn and Toomey (2001)



upper crust	passive seismic / S-waves	8 - 30 %	Fast direction parallel to the axial valley	Barclay and Toomey (2003)
upper crust	active seismic	2 – 3 %	Fast direction aligns with ridge axis	Dunn et al. (2005)
upper and intermediate crust	passive seismic	-	Fast direction parallel to the spreading axis	Bacon et al. (2022)
upper and intermediate crust	active seismic	1.5 – 5.2 %	Fast direction aligns with rift axis	Kidiwela et al. (2026)

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206 The magnitude of azimuthal anisotropy is estimated from the peak-to-peak traveltime
 207 variation relative to the representative propagation time through the sampled crust. The traveltime
 208 through the crust was determined by averaging the individual traveltimes within each ray group and
 209 subtracting the traveltime through the water column. Using the present analysis, the shallowest ray
 210 group yields an apparent anisotropy of ~4.5 - 5.5%, whereas the group with intermediate penetration
 211 depths of ~1 - 2 km yields ~3 - 4 %. These values should be interpreted as effective estimates for
 212 cumulative ray paths rather than as independent measurements of discrete 0 - 1 km and 1 - 2 km
 213 layers, because deeper rays also traverse the overlying crust. For rays penetrating into the lower crust
 214 down to ~2 - 3 km below seafloor, substantial residual variability remains, but no coherent twofold
 215 azimuthal component is resolved. We therefore do not infer that anisotropy is absent at these depths;
 216 rather, it diminishes with depth, either because faults may not reach these depths or because they are
 217 sealed much faster than shallow faults due to the increase in confining pressure. In addition,
 218 anisotropy could be more complex due to the heterogeneous gabbroic architecture of the lower oceanic
 219 crust. The observed 1 Θ -pattern might partly also a result of a lateral velocity trend, which is not
 220 considered in the initial isotropic velocity model and could result in the observed pattern. Assuming a
 221 lateral velocity trend, the velocities in the W would be slightly higher than in the E.

222 The degree of anisotropy observed in this study compares well to observations by other
 223 authors: Furumoto et al. (1971), White and Whitmarsh (1984), Shearer and Orcutt (1985) and Dunn
 224 and Toomey (2001) all observed anisotropy between 0 % and 7 % within the upper crust along the
 225 Hawaiian Ridge, across the Mid-Atlantic Ridge, within the Pacific Plate and across the East Pacific
 226 Rise, respectively. Higher degrees of anisotropy between 8 % and 30 % were found by Stephen (1985)
 227 and Barclay and Toomey (2003); however, they constrained these high degrees of anisotropy to the
 228 uppermost 500 m of the crust, which is generally considered one of the most heterogeneous parts of
 229 the shallow lithosphere. In addition, Barclay and Toomey (2003) analyzed S-wave data, which are
 230 known to yield significantly higher degrees of anisotropy (Alkhimenkov et al., 2020). Previous studies
 231 analyzing anisotropy at spreading centers observed the highest degree of anisotropy at shallow depths
 232 and a decrease with increasing depth. They suggested that the crack and fracture density is highest at



233 shallow depths and diminishes with increasing depth due to increasing pressure and progressive
234 sealing and alteration (Stephen, 1985; Dunn and Toomey, 2001; Barclay and Toomey, 2003). Even
235 though the anisotropy in this study is not aligned along the spreading axis and the Nazca plate in our
236 study area is much older than in most of the other studies, we assume similar mechanisms to explain
237 the observed reduction in anisotropy with depth.

238 **5.2 Resolution and uncertainty analysis**

239 To assess the resolution and significance of our results, we have conducted two synthetic
240 forward modeling tests along an E-W oriented 2D profile crossing the acquisition circles. These tests
241 were performed using the *modeling* interface (Fujie et al., 2008) for the RAYINVR code from Zelt
242 (1999). The first test aimed to investigate whether we can resolve anisotropy using our acquisition set-
243 up. Therefore, we modeled traveltimes using a velocity model adapted from Warwel et al. (2025b) as
244 well as using a velocity model that was increased by the observed degree of anisotropy. The resulting
245 traveltime differences are between 30 ms and 100 ms, depending on the offset (Fig. 4a). For the
246 shallow and the intermediate depth interval, the residuals are 30 – 50 ms and 50 – 80 ms, respectively.
247 These values demonstrate that velocity perturbations of the inferred magnitude can produce traveltime
248 differences comparable to the observed azimuthal residual amplitudes. However, because this test
249 changes velocity along a 2D profile rather than simulating the full circular acquisition through an
250 explicitly anisotropic medium, it should be regarded as a sensitivity test rather than a complete
251 recovery test for azimuthal anisotropy or its vertical distribution. In the second test, three 150-m-wide
252 vertical low-velocity fault zones extending to ~2.5 km depth were added to the model, with velocity
253 reductions of 20% at shallow depths and 10% near the fault bases (cf. Roland et al., 2012). The
254 resulting traveltime perturbations reach ~50 ms (Fig. 4b), showing that discrete fault-zone
255 heterogeneity can produce residuals of the same order as the observations. This test therefore supports
256 a structural origin for the signal but also emphasizes that effective anisotropy and spatially organized
257 isotropic heterogeneity are alternative, and potentially complementary, descriptions at the scale of the
258 experiment.

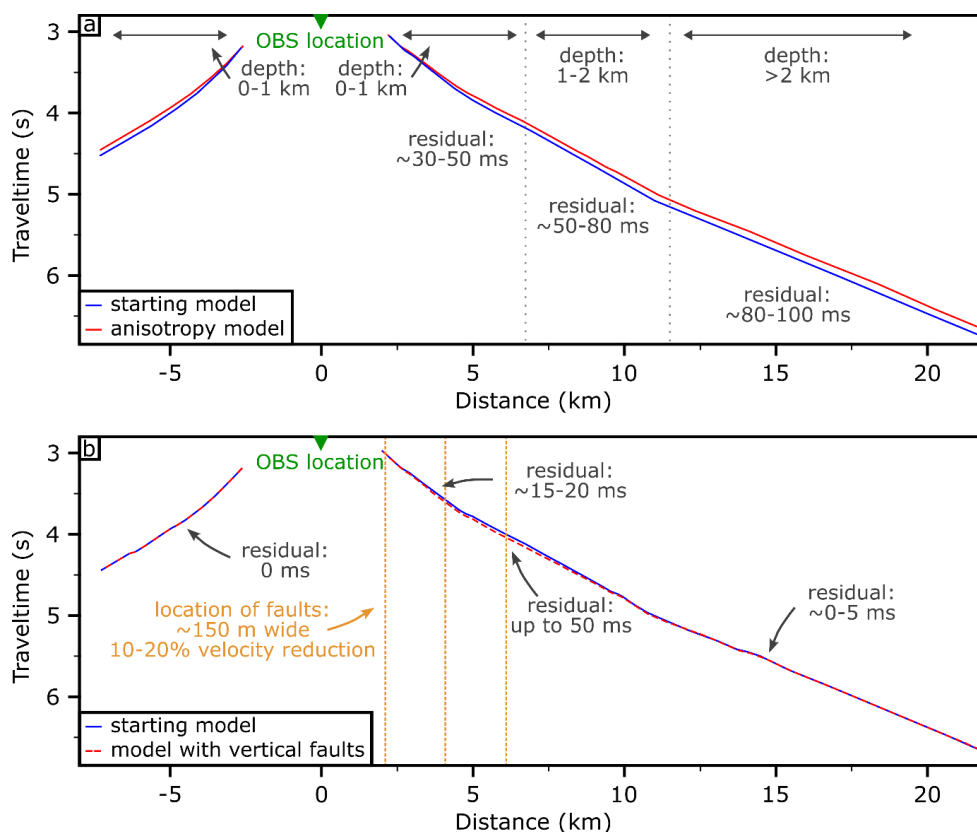


Figure 4: a) Synthetic traveltimes for OBH102 through a 1D velocity model based on Warwel et al. (2025b) (blue) (extracted at 20 km of Fig. 4 of Warwel et al. (2025b)) and added anisotropy from this study (red). b) Synthetic traveltimes for OBH102 through a 2D velocity model based on Warwel et al. (2025b) (blue) and added vertical faults (dashed red).

Even though this modeling shows that we can resolve crustal anisotropy, we have to account for a number of uncertainties in data acquisition and processing. Regarding the acquisition, we have to account for uncertainties in the locations of the OBS/OBH on the seafloor, as well as for uncertainties in the source location and non-linear internal clock time drifts. On the processing side, uncertainties may arise from picking errors, limitations of the forward modeling and fitting cosine curves to the data. The main causes of uncertainties, as well as their quantification or an estimate of their impact, are listed in Table 2. The locations of the OBS/OBH on the seafloor are obtained through triangulation using the direct water wave. Subsequently, the locations are verified by applying a static correction to the direct water wave, which yields a flat direct wave across all traces/offsets when the perfect location is found (Fig. S3). As the direct wave only travels through the water column, an accurate water velocity is essential. Based on a sound velocity profile, we determined an average water velocity of 1499 m/s. Remaining undulations in the water wave signal indicate an inaccurate location or depth of



the station, and the amplitude of these undulations can be used to assess their impact on the data. For OBS101, OBH102 and OBS103, we observe undulations in the range of 10 ms or less, while for OBS100 and OBS104, the undulations are in the range of ~15 ms (Fig. S3). In all cases, the resulting error is smaller than the peak-to-peak difference in the cosine curves and therefore does not overprint the crustal anisotropy. Source positions were determined from a GPS antenna mounted on a floating platform above the G-gun array. Position errors are expected to be largely random or to produce a fixed spatial bias (1 Θ -pattern) rather than a coherent twofold 2 Θ -pattern centered on each OBS/OBH. The motion of the array relative to the antenna due to currents and waves is estimated to contribute errors of only a few milliseconds.

Table 2: Summary of the main factors of uncertainty/errors, their quantification or estimate, as well as our interpretation of their impact.

Uncertainty / Error	Quantification / Estimate	Impact
OBS/OBH location	101/102/103: ≤ 10 ms; 100/104: ~15 ms	Minor impact
Source location	-	Not significant
Pg picks	25 – 50 ms	Potential effects on individual data points, but limited impact on the overall pattern
Ray tracing	-	Not significant
Velocity trend not considered in isotropic model	-	Not significant for upper crust; Potential impact for lower crust (1 Θ -pattern)
Fitting cosine curves	< 10° striking direction; <10 ms peak to peak difference	Not significant

The largest individual-data uncertainty is associated with picking the refracted Pg first arrivals. The signal length (time from the beginning of the phase to its end) varies between ~50 ms for the small acquisition circle and ~100 ms for the largest circle (Fig. S1). This results in theoretical picking uncertainties between 25 ms and 50 ms, respectively. This is only slightly smaller than the observed residual variations and therefore poses the threat of misinterpretation. However, even more important than picking the correct arrival time is picking a consistent phase. As visual inspection of the picks suggests that the picks follow the same phase (Fig. S1), the actual picking error is estimated to be lower than 25 - 50 ms. In addition, the azimuth-dependent residuals in Figure 3 result from averaging picks from different stations. The impact of individual picking errors is therefore assumed to average out. We assume that picking errors may impact individual data points but not the overall residual pattern. We apply a similar argument to rule out a misinterpretation arising from insufficient sampling frequency. The data were recorded at 250 Hz, implying that 1 data sample has a length of 4 ms. The observed peak-to-peak difference for the shallowest depth interval is 58 ms, and thus, is represented by only ~15 data samples. As the data points for each azimuth used to fit the cosine curves



302 represent averaged residuals from different stations and shots, we assume that this is sufficient to
 303 interpret the cosine curves.

304 Ray tracing performed by the *TOMO3D* code is based on a combination of the graph and the
 305 bending methods, which are both based on Fermat's principle (Meléndez et al., 2013, 2015).
 306 Inaccuracies of the ray tracing may affect our investigation, as we assume that traveltimes residuals
 307 result from velocity variations relative to the isotropic starting model. However, significantly different
 308 ray paths compared to the actual seismic ray propagation could also result in traveltimes residuals.
 309 Since the graph and bending methods, as well as the *TOMO3D* code, are well established and widely
 310 used for the inversion of seismic data, we assume that ray tracing does not significantly affect our
 311 results. However, the isotropic starting model itself could be a source of uncertainty, as it does not
 312 include potential lateral velocity variations that are unrelated to anisotropy and that can lead to pattern
 313 such as the 1Θ -pattern observed in the lower crust.

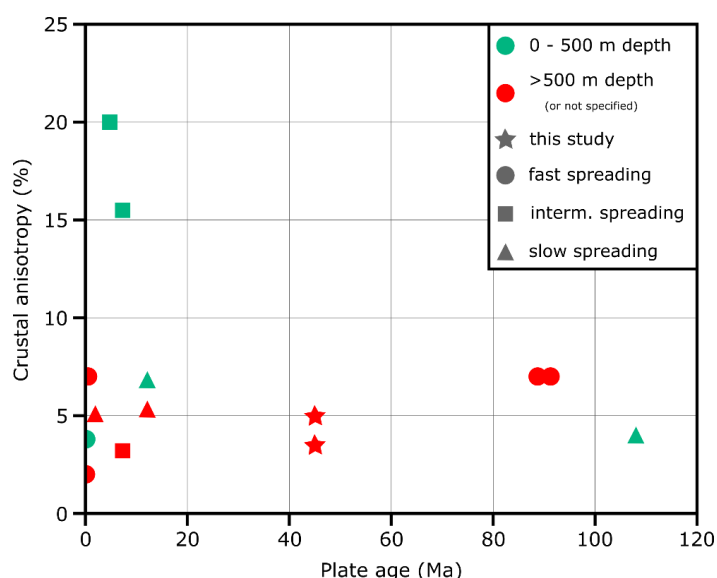
314 The uncertainty introduced by fitting a cosine curve to the data can be assessed and quantified
 315 by analyzing the standard error of each cosine curve parameter. We calculated cosine curves from the
 316 maximum and minimum values of each parameter and derived their azimuths of maximum and
 317 minimum, as well as their peak-to-peak difference (Fig. S2). This analysis shows that the direction of
 318 the interpreted fast and slow symmetry axes varies by less than 10° . The peak-to-peak residual
 319 difference varies between 54 ms and 62 ms for the shallowest depth interval and 74 ms and 81 ms for
 320 the intermediate depth interval. This small variation does not affect the estimated degree of anisotropy
 321 for the shallowest and intermediate intervals. In general, minor variations in the striking direction are
 322 possible, but uncertainties in the cosine curves do not impact the general alignment with the horst and
 323 graben structures. The same holds for the degree of anisotropy as well as its decay with depth.

324 5.3 Factors impacting crustal anisotropy

325 The influence of plate age and spreading rate on oceanic crust anisotropy has not been
 326 systematically established because the available observations are sparse and methodologically
 327 heterogeneous. Figure 5 compiles published anisotropy estimates as a function of spreading rate and
 328 plate age. The largest reported values ($\sim 15 - 20\%$) occur in crust younger than ~ 10 Ma and were
 329 measured at very shallow depths (< 500 m), where fracture density and structural heterogeneity are
 330 expected to be greatest (Fig. 5) (Stephen, 1985; Iturrino et al., 2005a; Furumoto et al., 1971; White
 331 and Whitmarsh, 1984; Shearer and Orcutt, 1985; Dunn and Toomey, 2001; Barclay and Toomey,
 332 2003; Iturrino et al., 2005b; Dunn, 2015). In general, slow spreading rates are associated with tectonic
 333 faulting, whereas fast spreading rates are associated with axial highs and reduced faulting (e.g., Phipps
 334 Morgan and Chen, 1993; Buck et al., 2005). As anisotropy is commonly related to faults created at
 335 spreading centers, an anisotropy-dependence of the spreading rates would be conceivable. However,
 336 the compiled data do not support this idea, as most of the anisotropy values for fast, intermediate and
 337 slow spreading rates plot within similar ranges (Fig. 5). Grevemeyer and Weigel (1996) observed a



338 significant age-dependent crustal velocity increase for crust younger than 10 Ma, while for older crust
 339 they did not observe significant velocity variations. Regarding crustal seismic anisotropy, a
 340 dependence on crustal age is not discernible based on the available data. The degrees of anisotropy
 341 observed at plate ages of >40 Ma are generally within the same range of anisotropy as for young plates
 342 (<15 Ma) (Fig. 5). However, the sparse distribution of anisotropy observations impedes a
 343 comprehensive analysis of anisotropy variations with increasing crustal age. The highest degrees of
 344 anisotropy are observed within the shallow crust of less than 500 m. While the compilation shown in
 345 Figure 5 itself is not conclusive, multiple studies, including our results, have observed a depth
 346 dependency of crustal anisotropy (e.g., Dunn and Toomey, 2001; Barclay and Toomey, 2003; Dunn,
 347 2015). We conclude that the sampling depth and the local tectonic setting are important controls on
 348 crustal anisotropy, while the independent effect of plate age and spreading rate remains unresolved by
 349 the current global dataset.



350

351 **Figure 5:** Compilation of anisotropies in different oceanic crusts and for different plate ages. Data are
 352 compiled after Furumoto et al. (1971), White and Whitmarsh (1984), Shearer and Orcutt (1985),
 353 Stephen (1985), Dunn and Toomey (2001), Iturrino et al. (2005) and Dunn (2015). If the age of the
 354 plate and/or the spreading rate is not given in the respective study, it is estimated from Müller et al.
 355 (2008).

356 5 Conclusion

357 We present active-source constraints on azimuthal P-wave anisotropy in a ~45 Ma old
 358 segment of the incoming Nazca Plate offshore Chile (~25.4°S), where direct offshore observations of
 359 crustal anisotropy have previously been lacking. Pg traveltimes residuals from a circular acquisition



show a coherent and systematic $\cos(2\Theta)$ dependent, twofold azimuthal pattern for rays sampling the upper ~ 2 km of crust, with an approximately E-W fast direction and N-S slow direction. The fast direction is approximately parallel to younger E-W horst-and-graben structures related to N-S extension and is oblique to the inferred NW-SE paleo-spreading fabric. This geometry indicates that younger intraplate deformation can dominate and potentially overprint the anisotropic fabric inherited during crustal accretion. The present analysis yields apparent anisotropy of up to 5.5 % for shallow-penetrating rays and a weaker estimate for rays penetrating deeper. For rays reaching below ~ 2 km into the lower crust, no coherent anisotropic signature is resolved; this does not require the deeper crust to be isotropic but places any anisotropic contribution below the present resolution or indicates that it is obscured by lateral heterogeneity. The weakening of the coherent signal with increasing penetration depth is consistent with decreasing fracture compliance and/or density under increasing confining pressure and progressive alteration or mineral sealing. Comparison with previous studies further indicates that sampling depth and local tectonic fabric are major controls on measured oceanic-crust anisotropy, whereas the independent roles of plate age and spreading rate remain insufficiently constrained. These observations highlight the importance of post-spreading fault architecture for the structural and hydrological preconditioning of oceanic crust before subduction.

Data Availability

The active-source seismic data and the high-resolution bathymetry data are available in the Zenodo data repository (Warwel et al., 2025c, <https://doi.org/10.5281/zenodo.16361774>; Warwel et al., 2026, <https://doi.org/10.5281/zenodo.22217214>).

Author contributions

DL, AD, MR and HK designed the anisotropy experiment during SO297. AW processed the OBS data, picked the traveltimes, created the figures and drafted the first version of the manuscript. AD conducted the forward modeling. AW, AD, AJ, MR, DL and HK were involved in the interpretation and the design of the manuscript. HK and DL provided the project management and funding acquisition. All authors contributed to writing and editing the manuscript.

Competing interests

The authors declare that they have no conflict of interest.

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