Radial anisotropy and S-wave velocity depict the internal to external zones transition within the Variscan orogen (NW Iberia)

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Abstract. The cross-correlation of ambient noise records registered by seismic networks has proven to be a valuable tool to obtain new insights into the crustal structure at different scales. Based on 2- to 14-s-period Rayleigh and Love dispersion
10 data extracted from the seismic ambient noise recorded by 20 three-component broadband stations belonging to two different temporary experiments, we present the first i) upper crustal (1-12 km) high-resolution shear wave velocity and ii) radial anisotropy variation models of the continental crust in NW Iberia. The area of study represents one of the best exposed cross-sections along the Variscan orogen of western Europe, showing the transition between the external eastern zones towards the internal areas in the west. Both the 2-D maps and an E-W transect reveal a close correspondence with the main
15 geological domains of the Variscan orogen. The foreland-fold and thrust-belt of the orogen, the Cantabrian Zone, is revealed by a zone of relatively low shear wave velocities (2.3-3.0 km/s), while the internal zones generally display higher homogeneous velocities (> 3.1km/s). The boundary between both zones is clearly delineated in the models, depicting the arcuate shape of the orogenic belt. The velocity patterns also reveal variations of the elastic properties of the upper crust that can be linked to major Variscan structures, such as the basal detachment of the Cantabrian Zone or the stack of nappes
20 involving pre-Variscan basement; or sedimentary features such as the presence of thick syn-orogenic siliciclastic wedges. Overall, the radial anisotropy magnitude varies between -5 and 15 % and increases with depth. The depth pattern suggests that the alignment of cracks is the main source of anisotropy at < 8 km depths, although the intrinsic anisotropy seems to be significant in the West-Asturian Leonese Zone, the low-grade slate belt adjacent to the Cantabrian Zone. At depths > 8 km, widespread high and positive radial anisotropies are observed, which we attribute to the presence of subhorizontal
25 alignments of grains and minerals in relation to the pre- or syn-orogenic deformation associated with the Variscan orogenesis.

1 Introduction

Seismic anisotropy is a ubiquitous feature within the Earth's interior that provides valuable information about the fabric of the geological materials and the tectonic and geodynamic processes causing it (e.g. Silver, 1996). It is usually investigated
30 through laboratory tests on rock samples (e.g. Mainprice *et al.*, 2000; Godfrey *et al.*, 2000; Ji *et al.*, 2015) or estimated

indirectly from the direction-dependent velocity variation of the seismic waves travelling across a rock formation (e.g., Luo *et al.*, 2013; Moschetti *et al.*, 2010). Vertically (V_{sv}) and horizontally (V_{sh}) polarized shear wave velocities can be extracted from Rayleigh and Love dispersion measurements, respectively. When $V_{sh} > V_{sv}$, the anisotropy is considered to be positive, whereas when $V_{sv} > V_{sh}$ the anisotropy is negative. Within an isotropic medium, V_{sv} and V_{sh} are equal. However, numerous studies in several locations worldwide have observed the so-called 'Rayleigh-Love discrepancy' (Anderson, 1961), which implies that the Rayleigh and Love wave dispersion curves cannot be simultaneously fitted under the assumption of isotropic wave velocities. Following Anderson (1961), this mismatch indicates that the medium is radially anisotropic.

Crustal radial anisotropy (RA) is originated by a variety of source mechanisms that dominate at different depth levels. In the shallow crust (<5-10 km) the RA is believed to be caused by networks of cracks and fractures linked to the regional stress-field (e.g. Crampin *et al.*, 2015). The presence of structure-oriented cracks, bedding planes or grain boundary alignments within the rockmass (e.g. Almqvist and Mainprice, 2017; Boness and Zoback, 2006; Zinke and Zoback, 2000) can also play an important role in the anisotropy from shape-preferred orientation or mineral grains. At pressures higher than 150-200 MPa, equivalent to 6-8 km depths, the increasing lithostatic pressure forces the closure of the cracks and texture-related anisotropy becomes dominant (e.g. Ji *et al.*, 2015; Kern, 1990; Christensen 1985). This type of anisotropy arises from the alignment of minerals with a crystallographic preferred orientation, due to deformation associated to faults and shear zones (Barruol and Kern, 1996). Thus, the effect of this progressive shift in the anisotropy-governing features is expected to be observed by geophysical methods targeting the upper 10 km of the crust, considering that the shallowest part is sensitive to the properties of the fractures within the rock massif, while the deepest sectors are more susceptible to mineral orientation in the rocks. The strongest crustal shear wave anisotropies, up to 20%, have been reported in recent orogenic belts, such as the Himalayas (e.g., Xie *et al.*, 2013) or the Andean Cordillera (Lynner *et al.*, 2018). Nonetheless, high RAs (10-16%) have also been identified in many other parts of the world with very different tectonic histories, even in sedimentary basins (Dreiling *et al.*, 2018), orogenic belts (Wang *et al.*, 2020) or areas with intra-plate volcanism (Ojo *et al.*, 2017). Besides, lateral variations in the crustal anisotropy have been related to a wide catalogue of geological features: microcracks and metamorphic foliations (Guo *et al.* 2012), fault systems (e.g. Shirzad and Shomali, 2014; Shirzad *et al.*, 2017), geological boundaries (e.g., Naghavi *et al.*, 2018; Wang *et al.*, 2020) or lateral crustal flow (e.g. Lynner *et al.*, 2018).

In this work, the first 3D crustal radial anisotropy model of NW Iberia is derived from the inversion of surface wave group velocities obtained from ambient noise interferometry (ANI) (e.g., Wapenaar, 2003, 2004; Shapiro and Campillo, 2004; Shapiro *et al.*, 2004; Snieder, 2004). This area (Fig. 1a) is a geologically complex region with a long and well-established tectonic history that comprises the overlapping structural imprint of two major compressional events: the Variscan orogeny, in the Late Paleozoic; and the Alpine orogeny, in Cenozoic times, separated by several rifting episodes in the Mesozoic. The structure of the study area was mainly configured during the Variscan orogeny, although the Alpine compression reworked some of the structures and originated a widespread pervasive fracturing, particularly in the upper crust. Overall, the part of the Variscan belt that crops out in the Cantabrian Mountains (CM) represents one of the most complete sections of this orogen in Europe (Martínez-Catalán *et al.*, 2009), illustrating the transition from the internal (West)

65   to the external (East) zones of an orogenic system (Fig. 1b). From the early 90's, extensive deep seismic reflection and refraction/wide angle surveys were carried out in this area. These programs focused in the unveiling of the deep crustal structure of the CM (Pérez-Estaún *et al.*, 1994; Pulgar *et al.*, 1995, 1996; Álvarez-Marrón *et al.*, 1996, 1997; Gallastegui *et al.*, 1997, 2002; Ayarza *et al.*, 1998; Fernández-Viejo *et al.*, 1998, 2000; Fernández-Viejo and Gallastegui, 2005; Fernández-Viejo *et al.*, 2011 2012) and broadened the knowledge about the multiorogenic crustal architecture. To improve imaging

70   resolution at crustal depths, several temporary seismic arrays with different specifications and dimensions were deployed in the CM (e.g. López-Fernández *et al.*, 2012; 2018), the most recent of which is featured for the first time in this study.

     Similarly to other passive seismic imaging techniques, the ANI reflects the variation of the seismic velocities of the bulk rock. In orogenic belts, where lithologies with different properties are placed in contact by structural features, ANI can also help to infer indirectly the presence of structures at depth and thus it can be valuable for unravelling the upper crustal

75   structure at regional or even local scale in a cost-effective manner (e.g Zigone *et al.*, 2015; Sammarco *et al.*, 2017; Acevedo *et al.*, 2019; Gu *et al.*, 2019; Olivar-Castaño *et al.*, 2020). In any case, the main advantage of ANI is that, unlike earthquake-based tomography, it can be applied in seismically quiet regions due to the ubiquitous presence of noise sources. Up to the date, there is limited knowledge on the seismic anisotropy signature of NW Iberia. Díaz *et al.* (2002, 2006, 2015) studied the upper mantle and lower crust anisotropy from distant earthquake shear wave splitting measurements and Acevedo *et al.*

80   (2020) used local earthquake shear wave splitting and ANI analysis to infer the anisotropic properties of the upper crust in the central part of the CM. Both studies consistently reported an average E-W fast direction, but the rotation of the fast orientations in the vicinity of major Alpine structures suggest that the contribution of the Alpine deformation to the anisotropy in the upper crust can be significant (Acevedo *et al.*, 2020).

     Here we make use of the ambient noise recordings from a new temporary seismic network in NW Iberia to extend to the

85   North and the West an earlier tomographic S-wave velocity model of the eastern sector of the CM (Acevedo *et al.*, 2019), which was mostly affected by the Alpine deformation, and to further determine the 3-D distribution of the crustal radial anisotropy. From a subsequent work (Acevedo *et al.*, 2020), we learnt that the crustal fabric controlled the orientation of fastest seismic velocities, sometimes in accordance with Variscan structures, others with Alpine structures. However, there was also local coincidence in the orientation of Variscan and Alpine structures, mostly by reworking during the youngest

90   convergence. The aim of the new seismic array, building on our previous experience, was two-fold. Firstly, to apply the ANI method to learn differences in architecture in relation to two orogenic events with contrasting kinematics (therefore orientation of structures). For this reason, we extended the area of observations to include parts of the Iberian crust with stronger Variscan imprint, incorporating domains with strong tectonic fabric susceptible to produce seismic anisotropy (Cárdenes et al., 2021) and with contrasting orientation to Alpine structures (Figs. 1a and b). Secondly, to constrain in the

95   upper crust the contribution to seismic anisotropy by internally deformed rocks versus tectonically fractured rocks.

2 Geological setting

2.1 Tectonic history

The crustal architecture of the study area is mainly defined by the Variscan orogeny, which affected the continental crust of western Europe and the northwest of Africa between the Late Devonian and the Early Permian (e.g. Matte *et al.*, 1986; Ribeiro *et al.*, 2007; Martínez-Catalán *et al.*, 2009; Pastor-Galán *et al.*, 2013). Subsequently to the denudation of the Variscan orogen, several rifting episodes affected the crust in NW Iberia, first in the Permian-Triassic, then in the Mesozoic where the crust was completely dismembered with the final opening of the Atlantic. The most important extensional episode occurred between the Late Jurassic and the early Cretaceous and was related to the opening of the Bay of Biscay (e.g. Cadenas *et al.*, 2018; Tugend *et al.*, 2015). This event configured the crustal structure of the eastern part of the study area, beyond the Ventaniella fault (Fig. 1a).

In Cenozoic times, the dominant geodynamic regime of the region shifted again and caused the North-South convergence between Europe and Africa, producing the onset of the Alpine orogeny. The compression started along the northern edge of the Iberian Peninsula in the Pyrenees and migrated to the West to form the CM (e.g. Teixell et al., 2018). During the convergence, the transitional crust of the Bay of Biscay was underthrusted southwards beneath the Iberian continental crust (e.g. Boillot *et al.*, 1979; Fernández-Viejo *et al.*, 2012; Cadenas and Fernández-Viejo; 2017). The bulk of the deformation was mainly accommodated by East-West oriented reverse faults and thrusts, but there are numerous examples of the reworking and retightening of favourably oriented earlier Variscan thrusts and folds, with a different degree of development across the study area (e.g. Alonso *et al.*, 1996). Currently, in the North Iberian crust, Alpine structures are dominant in the eastern half of the Variscan foreland-fold and thrust-belt, but they decrease their importance towards the West (Alonso *et al.*, 1996; Martín-González *et al.*, 2012; Llana-Fúnez and López-Fernández, 2015).

2.2 Crustal structure

Numerous works have contributed to broaden the knowledge about the Variscan imprint in the crust prior to the rise of the CM (e.g. Julivert, 1971; Pérez-Estaún *et al.*, 1988; Alonso *et al.*, 2009) and, more recently, its evolution during the Alpine convergence (Alonso *et al.*, 1996; Pulgar *et al.*, 1996, 1999; Gallastegui *et al.*, 2002, Martín-González *et al.*, 2012; Pedreira *et al.*, 2015). From the early 90's, extensive seismic surveys were carried out in the area. These programs focused in the unveiling of the deep crustal structure of the CM from deep seismic reflection and refraction/wide angle data (Pérez-Estaún *et al.*, 1994; Pulgar *et al.*, 1995, 1996; Álvarez-Marrón *et al.*, 1996, 1997; Gallastegui *et al.*, 1997; Ayarza *et al.*, 1998; Fernández-Viejo *et al.*, 1998, 2000; Gallastegui *et al.*, 2002; Fernández-Viejo and Gallastegui, 2005; Fernández-Viejo *et al.*, 2011, 2012) and led to significant findings that improved decisively the geological and geodynamical models of the area. For example, the E-W oriented profiles ESCIN-1 and ESCIN-3 imaged for the first time the Variscan structures of the crust under the CM and their progressive loss of importance compared to the Alpine structures towards the East. Later, the ESCIN-2 profile was performed with the goal of sampling the Alpine features in the transition between the eastern sector of

the CM and the Cenozoic Duero basin. Both profiles revealed the presence of an important crustal root beneath the CM. In order to complement the active geophysical surveys, several temporary seismic arrays with different specifications and coverage were deployed in the CM. Besides the seismotectonic characterization of the region (e.g. López-Fernández *et al.*, 2012, 2018; Llana-Fúnez and López-Fernández, 2015) the recorded passive seismic data allowed the study of the crustal configuration of North Iberia using receiver function analysis (Díaz *et al.*, 2003, 2009a; Mancilla and Díaz, 2015) and validated many of the ESCIN observations. The use of ambient noise interferometry is also beginning to provide additional knowledge about the internal structure of the upper crust in the study area (e.g. Acevedo *et al.*, 2019, 2020; Olivar-Castaño *et al.*, 2020).

2.3 Upper crustal domains in the study area

In the central sector of the CM the erosion of the Mesozoic sedimentary cover during the mountain building led to the exposure of the underlying Paleozoic basement. The physical properties of large parts of the upper crust targeted in this study with ambient noise tomography can be predefined according to the nature of the rocks and the structures, based on surface geology. As it has already been mentioned, a major part of the crustal architecture was shaped during the development of the Variscan orogeny and for that reason, the crustal domains that will be defined below follow partially the subdivision established for the Variscan orogeny (Julivert *et al.*, 1971). However, subsequent tectonic events, although did not produce a significant change in the nature of rocks, did have an impact on their structure. In most cases, producing a tendency to upright structures by increasing the dip of cartographic units and structures (e.g. Alonso, 1989).

The first crustal domain to the East of the study area is the foreland-fold-and-thrust belt of the Variscan orogen, the external part of the orogenic belt, corresponding to the Cantabrian Zone (CZ) in the literature (e.g. Alonso *et al.*, 2009). It is composed by sedimentary rocks that bear no tectonic or any other type of fabric orienting rock-forming minerals. However, the arrangement of tectonic units during the Variscan orogeny led in the final stages of the orogeny to the upright attitude of thrust sheets (e.g. Merino-Tomé *et al.*, 2009). Folding of syn-orogenic sediments not involved in earlier Variscan thrusting also led to the upright position of lithostratigraphic units (e.g. Alonso y Pulgar, 1989; Aller and Gallastegui, 1995).

To the West of the Variscan foreland-fold-and-thrust belt in the basement lies the hinterland of the orogen, showing an approximate N-S arrangement in the study area (Fig. 1a). In terms of the rock types, their fabric, and their attitude we could distinguish four domains: a linear slate belt wrapping around the external part of the Variscan orogen, a subhorizontal slate belt coinciding partly with the Mondoñedo nappe, a subvertical high grade metamorphic domain intruded by Variscan granitoids, and the allochthonous nappes of high-grade rocks sitting on top of the previous domain.

The linear slate belt is a domain that can be defined ahead of the Mondoñedo nappe, over the Navia-Alto Sil domain (Marcos, 1973). From the structural point of view, most of the structures are steeply dipping (Marcos, 1973) and they coincide with an area with a N-S lineation fabric (Matte, 1968). The rocks present a tectonic fabric developed at low-grade metamorphic conditions, which imposes a strong mostly planar mechanical anisotropy to the propagation of seismic waves (Cárdenes *et al.*, 2021). To the West, the subhorizontal slate belt coincides with the mid and rear part of the Mondoñedo

nappe (Fig. 1) and its structure is dominated by large recumbent folds (Matte, 1968; Bastida *et al.*, 1986). This isoclinal folding is accompanied by a tectonic fabric equilibrated in medium to high grade conditions that strengthens the subhorizontal anisotropy imposed by the orientation of the layers. Altogether, the linear and the subhorizontal slate belts constitute the West-Asturian Leonese Zone (WALZ).

165   The high-grade domain West of the Vivero fault (Fig. 1a) corresponds to the Central-Iberian Zone (CIZ). It is constituted by a rock sequence similar to that of the WALZ, but at higher grade given the pervasive intrusion of Variscan granitoids throughout the history of the orogenic belt. The recognition of large recumbent folds is hindered by the large amount of granitoids and by the development of upright folds and subvertical shear zones in the late stages of the Variscan orogen (e.g. Llana-Fúnez and Marcos, 2007). The folding is accompanied locally by the development of a strong crenulation

170 fabric (e.g. Bastida *et al.*, 2010). Subvertical corridors of foliated rocks in relation to large shear zones are common and often affect several kilometres on either side of the crustal scale structures (e.g. Llana-Fúnez and Marcos, 2001).

The last domain corresponds to the allochthonous nappes, which are thrust sheets of exotic provenance thrusted over the CIZ (e.g. Martínez-Catalán *et al.*, 1997). The presence of these allochthonous complexes defines the Galicia-Trás-os-Montes Zone (GTOMZ). Their essential features are that they are mostly made of high-grade high-pressure rocks, bearing a strong

175 tectonic fabric and showing a subhorizontal attitude (see e.g. Marcos *et al.*, 2002 for the Cabo Ortegal complex and Martínez Catalán, *et al.*, 2012 for the Órdenes complex).

The upper crustal domains presented have been defined by the types of rocks and structures formed during the Variscan orogeny. Subsequent tectonic events in the study area lack the intensity and the pervasive character shown during the Variscan orogeny. No tectonic fabric at regional scale has been reported in relation to either the several extensional episodes

180 leading to the opening of the Atlantic neither during the more recent Alpine convergence. Faults are also more spaced in between.

Earlier studies have reported in the CZ domain a somehow widespread effect of Alpine deformation over previous rocks and structures, dominated by the tightening of previously steep Variscan structures (e.g. Alonso and Pulgar, 1989). Locally in this domain, there are reports of evidence of post-Variscan fracturing. No systematic study of Alpine fracturing has been

185 carried out in the study area, however, separate observations in post-Permian rocks in the confines of the CZ show the development of subvertical N-S set of joints (e.g. Lepvrier and Martínez García, 1990; Pastor-Galán *et al.*, 2011; Uzkeda *et al.*, 2013).

3 Data and methods

3.1 Seismic data

190 This study used continuous broadband seismic data from two experiments with non-simultaneous recording periods: (1) Between June 2019 and February 2020, a seismic network (Geocantábrica-Costa Seismic Network, GEOCSN) composed of 11 portable stations was deployed in the western CM. The dataset from this network was complemented with data from two

permanent seismic stations belonging to the Spanish Seismological Network (SSN) of the Spanish Geographical Institute (Fig. 2a). The seismic array served the dual purpose of enhancing the location of the intraplate seismicity in the West of the

195   CM and the close continental shelf and acquiring new passive data for ambient noise analysis. The experiment covered an area of ~12,500 km^2 with interstation distances of ~40 km. Seismic stations were equipped with Nanometrics Taurus data loggers and broad-band Trillium 120 s sensors recording 3-component seismic data at a sample rate of 100 Hz. Most of the stations were installed in isolated chapels and sheds with connection to the electricity network, although solar power supply was required in some locations. A few stations were also mounted directly in the field, buried in ~0.5 m pits. The average

200   data recovery reached 96%, with minimum data loss due to occasional power cuts caused by thunderstorms. (2) To extend our observations to the West, we also processed 12 months (January-December 2012) of continuous seismic data registered by 7 temporary stations of the northern IberArray seismic network (Díaz *et al.*, 2009b) plus the station EPON of the SSN (Fig. 2a). The seismic recording equipment and their technical configuration were the same as in the GEOCSN, except that all the seismic stations were mounted outdoors. The station spacing was also larger, around 60 km, covering an area of

205   ~10,000 km^2 with a 26% overlap with the GEOCSN.

3.2 Ambient noise interferometry

The treatment of the seismic ambient noise data to obtain Rayleigh- and Love-wave dispersion velocities was based on the widely used processing scheme described by Bensen *et al.* (2007), with a few modifications. Our processing began with the splitting of the continuous dataset in 24-hour long segments and the removal of all the traces with gaps. The records were

210   downsampled to 25 Hz and the instrument responses were removed. Next, the mean and the trend were also removed from the traces and a band-pass filter with corner frequencies of 0.01 and 2.0 Hz was applied. The described pre-processing benefited from the use of the Phase Cross-Correlation technique (PCC, Schimmel, 1999) for the calculation of the cross correlations. Given that PCC is amplitude unbiased (Schimmel *et al.*, 2011), we did not perform any temporal or spectral normalization, thus avoiding the potential introduction of artifacts and shortening the processing time. The next step

215   consisted in the cross correlation of the 24-hour traces recorded simultaneously by each pair of stations with the aforementioned PCC procedure. In this way, we obtained the Daily Cross-Correlation Functions (DCCF) for lag times of 100 s. The observation of the cross-correlograms revealed a noticeable asymmetry between the causal and the acausal parts of the signal. Since we ignore which of these parts contain more useful information, the causal and the time reversed acausal parts of each station-pair DCCF were stacked using the Time-Frequency domain Phase-Weighted (tf-PWS) procedure of

220   Schimmel and Gallart (2007). This stacking method enhances the coherent signal and cancels incoherent noise, allowing the extraction of the symmetrical empirical Green's functions (EGF) of the medium between each station pair. In this study, we processed the vertical (Z) and the horizontal components (E, N) to obtain the Z-Z (Fig. 3a), E-E and N-N EGFs. Then, the horizontal components were rotated along the azimuth of the interstation path to provide the radial (R-R, Fig. 3b) and transverse (T-T, Fig. 3c) correlation functions. The vertical and the radial components are dominated by Rayleigh waves

225   (Figs. 2a and 2b), whereas Love waves emerged from the transverse component (Figs. 2a and 2c).

The measurement of the Rayleigh and Love group velocity dispersion curves was performed by using the multiple filter analysis (MFA) method of Dziewonski *et al.* (1969). In this technique, a narrow-band Gaussian filter is applied over different periods to isolate the wave package around the central period of each filter (Figs. 2d and S1). The group velocities in their fundamental mode were then manually picked to construct dispersion curves. According to Bensen *et al.* (2007),

230 dispersion curves are limited at the longer periods by the fact that velocity measurements are only reliable for interstation distances greater than three wavelengths. Given the dimensions of our network, the strict compliance of this criterion results in a loss of measurements that affects the resolution of the group velocity maps at periods > 10 s. However, Luo *et al.* (2015) demonstrated that consistent measurements can also be obtained using a cut-off of one wavelength.

In order to avoid discarding useful data without relaxing the constraints to a point at which the velocity determinations

235 become unreliable, we adopted a two-wavelength criterion in this work. This cut-off has been proven to be a suitable compromise solution in other studies (e.g. Sammarco *et al.*, 2017; Brandmayr *et al.*, 2016, Shapiro *et al.*, 2005). To ensure the quality of the results, our analysis was limited to velocity measurements within two standard deviations from the mean and with more than 60 measurements at each period. The stability of the dispersion curves was also investigated by stacking sets of 60, 90 and the available DCCFs for all the interstation paths (Fig. S2). Overall, Love wave dispersion curves are

240 slightly less stable than Rayleigh wave dispersion measurements. Waveforms and dispersion curves for the GEOCSN paths stabilize at a lower number of days (~60 days) than IberArray paths (~90 days) but, in both cases, the stabilization is reached at a much lower number of days than the available for each interstation path (Fig. S2).

We decided not to extract phase velocities due to the difficulty of resolving the 2-pi phase ambiguity without long period data, due to the small aperture of our seismic network and the complexity of the CM crust. The estimation of the

245 velocity uncertainties is vital to warrant the trustworthiness of the results. For this task, we have checked the group velocity mismatch between the dispersion curves obtained from 10 stacks made with the randomly selected 75% of all the available DCCF (Fig. S2). The results show, in general, uncertainties well below 1% for the Rayleigh and Love wave group velocity determinations from vertical and transversal components respectively. In relative terms, Love waves display higher velocity uncertainties than Rayleigh waves, so as the IberArray dataset when compared with the GEOCSN measurements. Compared

250 to vertical-component observations, Rayleigh wave velocities from radial-component EGFs display higher uncertainties, up to ~4% at the longest periods (Fig. S2), and a lower signal-to-noise ratio. These facts, combined with the loss of some interstation paths due to dubious radial MFA surfaces, drove us to use exclusively Rayleigh group velocities obtained from the high quality vertical EGFs in further calculations.

On the other hand, the stacking of noise-correlations over long time spans ensures that the results are not affected by a

255 seasonal distribution of the seismic sources. Moreover, the directionality analysis of noise sources in N Iberia performed by Olivar-Castaño *et al.* (2020) showed that, although the most intense noise comes from the N-NW directions (Bay of Biscay and Atlantic Ocean), a sufficient level of noise emerges from all azimuths. Therefore, the effect of the spatial variation of noise sources is small and does not compromise the reliability of the dispersion measurements.

3.2 Group velocity tomography and depth inversion

260   Reliable dispersion measurements of Rayleigh and Love waves were obtained at periods between 2-14 s (Figs. 2e and 2f). The transformation of the dispersion velocities between the stations into continuous Rayleigh and Love group velocity surfaces was performed with the Fast-Marching Surface Tomography scheme (Rawlinson, 2005). The FMST is an iterative non-linear method that is implemented in two steps: a) the forward prediction of travel-times with the fast-marching method (Sethian, 1996; Rawlinson and Sambridge, 2004a; Rawlinson and Sambridge, 2004b) and b) the modification of the model

265   parameters to explain the data observations, for which a subspace inversion scheme is used (Kennett *et al.*, 1988). The resulting models are controlled by a grid of nodes with cubic B-spline interpolation that generates a smooth and continuous velocity medium. We selected a grid size of 0.1° x 0.1° for the inversion of a homogeneous starting model that was created from the average velocity measured for a certain period (Nicolson *et al.*, 2014). After extensive tests, the smoothing and damping parameters were set to 0.001 and 0.1, respectively, but their change only had a minor influence in the resulting

270   models.

     Based on the set of tomographic group velocity maps, we created new pseudo-dispersion curves by extracting the Rayleigh and Love velocity values for each point of a 0.1° x 0.1° grid. The independent inversion of these curves with the CPS surface inversion codes (Herrmann, 2013) allowed us to reconstruct the variation of the V_{sv} and the V_{sh} with depth in each grid node. In order to address the non-uniqueness nature of the inversions, we calculated two initial regional S-wave

275   velocity model to be used as input in each of the subsequent computations. These models were obtained in a previous step from the separate inversion of the average Rayleigh and Love dispersion curves in the whole study area (Figs. 2e and 2f). We started in all cases from a homogeneous 22-layer model, the first four layers with a thickness of 0.5 km and the rest with a thickness of 1 km. Based on published studies in the area (Fernández-Viejo *et al.*, 2000), we set a S-wave velocity of 3.35 km/s for each layer, with a fixed V_p/V_s ratio of 1.74 and a density of 2,600 kg/m^3. The mismatch between the S-wave

280   estimations from Rayleigh and Love waves allowed us to assess the intensity and distribution of the RA following Eq. (1):

$$RA(\%) = \frac{2(V_{sh} - V_{sv})}{(V_{sh} + V_{sv})} 100 \; , \tag{1}$$

(Yanovskaya *et al.*, 1998; Savage, 1998; Ojo *et al.*, 2017). All the inversions were performed with a damping factor of 10 in the first iterations to avoid an excessive shift from the initial model, whereas a value of 0.5 was used for the remaining iterations.

285   **4 Results**

4.1 Rayleigh and Love wave group velocities

     In total, we extracted Rayleigh and Love dispersion measurements of 78 station pairs from the GEOCSN network and 28 additional station pairs from the IberArray network. The 2-D tomographic maps displaying examples at selected periods

of Rayleigh and Love group velocity variations can be observed in Fig. 4. Considering the Rayleigh waves (Fig. 4, left panels), the absolute group velocities range between 2.3 and 3.1 km/s. On the other hand, Love waves (Fig. 4, right panels) show slightly faster velocities, varying between 2.3 and 3.4 km/s. At all periods, both Rayleigh and Love derived maps image two clear blocks: a high velocity zone occupying the western half of the study area and a relative low velocity zone in the eastern half. Both anomalies are separated by a narrow C-shaped transition sector which follows the boundary between the CZ and the WALZ and the surface trace of the main structures. The western anomaly presents high and moderately stable group velocities in the entire period range. On the contrary, group velocities in the eastern anomaly zone raise constantly as the period increases. The most noticeable difference between Rayleigh and Love maps lies in the shape of the western anomaly at the highest periods. At those periods, Rayleigh group velocities display a light decrease, and the high velocities tend to concentrate in the South. However, Love group velocities delineate a consistent large high velocity surface in the western sector, while the low velocity anomaly migrates towards the South.

The resolution of the final velocity slices was investigated by performing synthetic checkerboard tests (Figs. S3 and S4). We used as reference three models with small (20 x 20 km), medium (30 x 30 km) and large (40 x 40 km) anomaly sizes and velocity perturbations of ±0.4 km/s. Then, we evaluated the performance of our array for recovering this previously known velocity structure at periods of 4, 8 and 12 s. For both Rayleigh and Love waves, at periods up to 8 s, the resolution is sufficient to resolve anomalies of ~20 km in the central and eastern sectors of the study area, and ~30 km in the westernmost zone. At periods longer than 8 s, the resolution decays due to the progressive decrease of the number of interstation paths and signs of smearing arise in the models. The average size of the recovered features increases to ~30 km. At those long periods, the Rayleigh paths recovered more accurate models than the Love paths. The RMS misfit between observed and predicted travel times ranged between 0.3-0.8 s, and 0.7-1.2 s for the Rayleigh and Love waves respectively, which indicates that the inverted models fit the observations and they are not affected by the selection of the regularization parameters (Nicolson *et al.*, 2012).

4.2 Shear wave velocities

Figure 5 shows examples of the S-wave inversion results in specific grid points located in the CZ (Fig. 5a), the WALZ (Fig. 5b) and the CIZ+GTOMZ (Fig. 5c), featuring a good fit between the observed and the inverted data. The models and the anisotropy characteristics from the WALZ and the CIZ+GTOMZ show similar patterns, whereas the CZ presents some peculiarities, such as a constantly increasing V_{sv} and V_{sh} velocity with depth and a pronounced negative radial anisotropy anomaly near the surface. Figure 6 shows examples of the variation of V_{sv} (Fig. 6, left panels) and V_{sh} (Fig. 6, right panels) at depths of 3, 6, 9 and 12 km. Within that depth range, the average V_{sv} increases from 3.0 to 3.4 km/s, while average V_{sh} oscillates between 3.1 and 3.5 km/s. Similarly to the group velocity maps, the presence of a low velocity anomaly to the East of the study area stands out both in V_{sv} and V_{sh} maps and its limit runs parallel to the curved boundary between the external and the internal zones, between the CZ and the WALZ. However, this anomaly attenuates at depth and finally disappears below ~6 km. At greater depths (>9 km) the correspondence between the features depicted in the V_{sv} and the V_{sh} maps

begins to fade away. In the V_{sv} surfaces, two new high velocity anomalies emerge in the South. However, the V_{sh} maps depict high velocities in the western/central sector and in the northeastern corner.

4.3 Radial anisotropy

325   The spatial distribution of the radial anisotropy (Fig. 7) was calculated from the discrepancies between V_{sv} and V_{sh} in each node of the tomographic grid (Figs. 5a, 5b and 5c). In absolute values, the whole study area shows a significant anisotropy that reaches 10-15% in certain areas. Nevertheless, there is a clear differentiation with depth. At depths shallower than ~8 km the anisotropy strength keeps within the ±6% range. Beneath those depths, the anisotropy starts to rise up to ~15%. Negative anomalies ($V_{sv} > V_{sh}$) are dominant at shallow depths in the eastern half and the southwestern limit of the

330   study area, with a weak positive corridor separating them. These negative anomalies become positive ($V_{sv} < V_{sh}$) at ~6 km and ~8 km depth, respectively.

In order to represent the velocity variations within the context of the Variscan structure of the CM, we constructed an E-W cross section that runs perpendicularly to the main Variscan elements (Fig. 8). The most striking feature of the V_{sv} (Fig. 8a) and the V_{sh} (Fig. 8b) cross sections is the low velocity anomaly located in the East. The V_{sh} section shows a more

335   extensive anomaly that coincides with the CZ. In the case of the V_{sv} section, the anomaly is more localized towards the sector where the syn-orogenic deposits of the Central Coal basin (CCB) are present. The previous S-wave tomography in Acevedo *et al.*, 2019 allowed us to extend the V_{sv} section until the transition between the CZ and the Basque Cantabrian basin (BCB). The higher velocities in both sections show a good correlation with the expected distribution of the Precambrian basement. In general, the internal zones show a more homogeneous velocity pattern, although V_{sv} is quite

340   constant and V_{sh} depicts a velocity increase with depth. The latter type of waves also register a slight reduction of velocity in the GTOMZ and CIZ domains in comparison with the WALZ. The distribution of anisotropy in the E-W section (Figures 8c and 9) also shows a strong correlation with the main geological domains (Figs. 1 and 8c). The mentioned positive corridor roughly coincides with the WALZ, while the negative patches are limited by the Vivero fault, that marks the beginning of the CIZ, and the Narcea antiform. To the East of the latter structure, the negative anisotropy is only observed in the part of

345   the CZ where syn-orogenic deposits are absent. On the other hand, the high positive anisotropies are roughly consistent with the presence of the lower Paleozoic and Precambrian basement rocks at depth.

5 Discussion

Tomographic maps in Figure 6 depict the first high-resolution S-wave models of the transition between the internal and the external parts of the Variscan orogen in NW Iberia. Considering the sensitivity of the method (Fig. S5), the discussion of

350   the results is restricted to the first 12 km of the crust. In addition, at the investigated periods (<14s) Love waves are primarily sensitive to the uppermost crust, while the sensitivity peak of Rayleigh waves deepens as the period increases. The following discussions are only focused on well-resolved areas from the checkerboard tests (Figs. S3 and S4).

5.1 Velocity structure in the upper crust

In previous ambient noise regional studies on the Iberian Peninsula, Macquet *et al.* (2014) and Palomeras *et al.* (2017) reported S-wave velocity values at 5 km depths ranging between 2.5 km/s and 3.5 km/s in the study area. Likewise, a recent work centered in the BCB also obtained this velocity range for the upper crust in the eastern termination of the CZ (Olivar-Castaño *et al.*, 2020). Silveira (2013) and Palomeras (2017) identified a first order anomaly which corresponds with the Iberian Massif and recognized some of the tectonic units that constitute it. However, due to resolution limitations, the transition between the internal and external zones was poorly constrained. In contrast, the most striking element in both surface-wave group and shear velocity models presented here is the precise delineation of the boundary between these domains by a sharp velocity change (Figs. 4 and 6).

The contour of this limit depicts a clear convex-to-the-west shape with the CZ located at its core. The contact between the high velocity anomaly, to the West, and the low velocity anomaly, to the East, is parallel to the Narcea antiform (NA), which is the structure that marks the boundary between the hinterland and the external zones of the Variscan orogen (Fig. 1) (Gutiérrez-Alonso, 1992; Pérez-Estaún *et al.*, 1994). The presence of rocks with completely different physical properties at both sides of the eastern boundary of the NA justifies the observed velocity contrast. The CZ represents the foreland fold and thrust belt of the Variscan orogen, constituted by a pre-orogenic sedimentary sequence deformed in a thin-skinned deformation context, and a syn-orogenic clastic sequence that was deposited during the formation of the orogen. On the other hand, the slate belt to the West, is a relatively homogeneous domain mainly composed by siliciclastic rocks affected by regional deformation fabric formed at low grade metamorphic conditions. In summary, the upper crust in the CZ is essentially made of sedimentary rocks without metamorphism while the upper crust in the NA and the WALZ slate belt is composed by igneous and volcanic rocks in the NA and low grade metamorphics towards the West, generating a clear contrast in the intrinsic physical properties of the rocks that are reflected in the crustal velocity patterns observed.

The sedimentary rocks of the CZ display relatively low S-wave velocities, typically under 3.0 km/s in the V_{sv} models or 3.2 km/s in the V_{sh} estimations. The profile of Figure 8a displays a good correspondence between the lower velocities and the location of the siliciclastic wedges of syn-orogenic sediments. In fact, the low velocity anomaly reaches its greater thickness, around 5 km, in the CCB, the largest of these wedges. The depth estimation in the CCB is consistent with previous interpretations (Aller and Gallastegui, 1995). Towards the West, both the V_{sv} and the V_{sh} models depict high velocities. V_{sv} remain constant about 3.3 km/s, while V_{sh} increases steadily with depth (Fig. 6). The rock sequences in the WALZ are quite homogeneous, although in the cross-section of Fig. 8a some high velocity patches can be identified between 5-9 km depth. The anomaly located close to the orogenic boundary is related with the stack and thrusting of rigid lower Paleozoic and pre-Variscan basement rocks from the NA (Pérez-Estaún *et al.*, 1994) and the exhumation of basement rocks, gneiss and pre-Variscan igneous rocks between the NA and the Allande fault (Marcos y Pulgar, 1982; Rubio-Ordóñez *et al.*, 2015). A smaller high velocity anomaly, in the West, is located near the Vivero fault (Fig. 8a) (López-Sánchez *et al.*, 2015). High velocities were expected in this area due to the presence of mafic rocks and Variscan granitoids. However, the V_{sh} profile

(Fig. 8b) registers a velocity reduction in this domain, maybe caused by the fact that the resolution in the edges of the models decays and velocities are less well constrained. The basement rocks at the surface also show pervasive fracturing and weathering, strongly affecting velocities.

5.2 Thickness of the sedimentary layer and basement depth

390   A significant result from S-wave models is the estimation of the thickness of the sedimentary layer. In Macquet *et al.* (2014), a S-wave velocity value of 2.9 km/s is proposed to indicate the limit between the basement and the sedimentary cover. In our case, this velocity defines quite well the position of the syn-orogenic sequence in the CZ, the presence of small Permo-Triassic basins in the CZ - BCB transition and the presence of Mesozoic sediments in the BCB but fails in recovering the depth of the basement. The main detachment of the CZ is believed to be located near the Precambrian-Cambrian

395   boundary (Pérez-Estaún *et al.*, 1988). Taking into account that the pre-Variscan basement rocks are exposed in the NA and show velocities around 3.1 km/s in Figure 8a, we considered that this can be a more appropriate value to estimate the depth of the basement-cover surface in our study area to infer the position of the detachment. The 3.1 km/s iso-velocity line initiates in the NA area. It stretches towards the East for 20 km at shallow depths and then it sinks almost vertically until reaching 8 km. This deepening coincides with the maximum eastern extension of the pre-Variscan rocks at depth.

400   Continuing to the East, the surface dips smoothly to the West and alternates ramps and flats until reaching depths of 1 km to the West of the Ventaniella fault. In the extension of Fig. 8a, based on the V_{sv} model of Acevedo *et al.* (2019), the iso-velocity line deepens again, showing an apparent dip towards the East. This change is related with the different structure of the CZ-BCB transition, which was mainly configured by the Alpine inversion of Mesozoic extensional faults (e.g. Espina, 1997; Tavani *et al.*, 2011). If we consider that the 3.1 km/s surface depicts the location of the base detachment of the CZ, its

405   greatest depth ranges around 8-9 km. This observation agrees with the studies of Pérez-Estaún *et al.* (1988, 1991, 1994), although it is slightly shallower than the estimations of Gallastegui *et al.* (1997, 2000).

5.3 Radial anisotropy

From the relation between V_{sv} and V_{sh} in Eq. 1 we have obtained the first model of RA in the continental upper crust in NW Iberia (Figs. 7, 8c, 9 and 10). Overall, the RA patterns display a fair correspondence with the main Variscan tectonic

410   units, at least in the shallowest crust (<7 km). At those depths, the distribution of the RA negative anomalies bound the GTOMZ-CIZ and the CZ, whereas in the WALZ a constant positive anisotropy is observed (Fig 8c). The sign of the RA is related with the inclination of the anisotropic features. Thereby a negative RA implies the presence of steeply dipping to subvertical structures (60-90°) while a positive RA suggest gently dipping to horizontal features (0-30°, Dreiling *et al.*, 2018; Xie *et al.*, 2013). Thus, in the CZ, the negative anomaly can be explained by the verticalization of Variscan nappes and other

415   structures, and the retightening of folds caused by the Alpine convergence, which was more intense in this domain than in any other. Moreover, the presence of a N-S oriented pervasive Alpine fracture system with subvertical dip has been recognized in the area and may play a significant role in the development of the anisotropy.

As the depth increases, the thrusts reduce their dip as they approach to their imbrication with the CZ basal detachment (Pérez-Estaún *et al.*, 1991), which is reflected in the progressive shift to a positive anomaly (Fig. 8c). There are two factors that suggest that this change can be provoked by the presence of a basal detachment at those depths. First, the positive RA points towards the presence of smoothly dipping anisotropic features. Second, the RA sign inversion occurs at greater depths beneath the western CZ than the eastern CZ, where the positive anomaly gets close to the surface. This point implies the presence of a possible structure with a general dip towards the West. Laterally, RA sign variations are also present, as in the transition between the CZ and the WALZ. From the Narcea antiform to the Mondoñedo thrust, subvertical structures are dominant, but the existence of a Variscan subhorizontal foliation parallel to the fold axis may compensate their effect and explain the slightly positive anomaly instead of the expected negative anisotropy. To the West of the MT, in the CIZ and the GTOMZ, the structures and foliation are mainly gently dipping, contrasting with the presence of a negative RA anomaly that spreads from the surface to 8 km depth. Strong subvertical deformation fabric related to late Variscan shear zones and folds (Llana-Fúnez and Marcos, 2007) and N-S fracture systems associated to the Alpine convergence have been reported in the area, but the resolution of the models at their edges is not optimal and further studies should be carried out to verify the existence of such negative anomaly.

In general, the magnitude of the RA is low, around 4%, at shallow depths (< 8 km). Considering the entire range of investigated depths, except for the WALZ, where the RA magnitude increases almost steadily, a progressive decrease followed by a rapid increment of the RA anisotropy magnitude is observed in the CIZ/GTOMZ and the CZ (Fig. 8c). The latter domain was also targeted by a previous ambient noise-based study that estimated the azimuthal anisotropy in the area (Acevedo *et al.*, 2020). A comparison between the RA measured in a representative CZ node and the azimuthal anisotropy determinations is displayed in Fig. 9a. Despite some discrepancies in their magnitude at shallow depths (< 3 km), both azimuthal and radial anisotropy depict a decline at depths ranging from ~3 to ~7 km. At greater depths, the RA rises gradually until it reaches magnitudes around 9% at 12 km, while azimuthal anisotropy increases abruptly at 7.5 km depth and stabilizes at 9%. The progressive closure of the open cracks due to the increment of the lithostatic pressure can explain the observed radial and azimuthal anisotropy magnitude reduction at shallow depths (Figs 9a and 9b). In crystalline low porosity rocks, the complete closure of microcracks at grain boundaries is observed in laboratory experiments at about ~200 MPa, which is equivalent to depths around 7 km (e.g. Christensen, 1985; Brown *et al.*, 2009; Guo *et al.*, 2014) and coincides with both the radial and the azimuthal anisotropy minimums (Fig. 9b).

At shallow depths (<5 – 10 km), the RA is believed to be caused by the alignment of stress- (Crampin, 1978; Crampin and Peacock, 2005) or structure-induced fluid saturated cracks (Zinke and Zoback, 2000). The fast anisotropic directions obtained from shear-wave splitting in Acevedo *et al.* (2020), which tend to align parallel to major structures, favor a structure-controlled anisotropy model for the CZ. What is more, although the Alpine deformation did not produce pervasive deformation in NW Iberia, it generated shallow fracture systems that can affect decisively the physical properties of the rocks near the surface, especially in the CZ and the CIZ/GTOMZ. On the other hand, in the case of slates from the WALZ, recent laboratory data have shown that the magnitude of the anisotropy increases with the decrease of confining pressure, as

the orientation of the microcracks is strongly controlled by the shape fabric in very well foliated slates (Cárdenes *et al.*, 2021). However, this means that another origin must be invoked to produce the RA magnitude increment at higher depths (Fig. 9a). As it is depicted in Fig. 1, the crystalline basement rock outcrops in the WALZ, showing an anisotropy that increases from 2% to 15% in depth. These Precambrian rocks, as well as the Lower-Paleozoic pre-Variscan siliciclastic sequences that form the WALZ and constitute the basement of the CZ, are affected by low-grade metamorphism and the development of a widespread tectonic foliation. Anisotropic minerals, such as phyllosilicates, are common in these rocks, so the alignment of minerals and grains (Wenk *et al.*, 2020; Cárdenes *et al.*, 2021) seems to be the most likely mechanism controlling the anisotropy in the study area and at depths > 7.5-8 km. In the CZ, this change in the deformation and tectonic history of the rocks controlling the anisotropy is also reflected in the fast azimuthal directions. At depths < 7.5 km, they remain steady at around 84°, whereas at depths at which the pre-Variscan basement gains influence in the results (> 7.5 km) a progressive NE rotation (~70°) can be noticed (Fig. 9a).

Many studies have tried to establish the linkage between tectonic regimes and RA patterns (e.g. Luo *et al.*, 2013; Dreilling *et al.*, 2018; Wang *et al.*, 2020). Generally, compressional and extensional regimes are associated to the vertical and horizontal alignment of cracks and minerals, respectively. However, the specific geodynamic setting of each location must be considered, given that both regimes can produce very different anisotropy signatures. The presence of negative RA at shallow depths in foreland fold-and-thrust belts has been observed in orogens worldwide. For example, Chen *et al.* (2009) and Guo *et al.* (2012) reported negative RAs in the Himalayan fold-thrust belt. Recently, Wang *et al.* (2020) described a widespread negative anomaly of ~5% below the Foreland belt of the Canadian Cordillera that turns into a positive anomaly in the Omineca Belt, the hinterland of the orogen. These authors suggest that the RA in the foreland fold and thrust belts is caused by the horizontal contraction of the upper crust, which creates subvertical faults and/or preferentially alignments of minerals under convergent regimes. The continuity of the negative anomaly at depths up to 20 km allowed them to infer a "thick-skinned" Cordillera-Craton transition. Contrarily, in our model, the negative anomaly in the CZ is distinctly limited at depth (Fig. 10), pointing to a "thin-skinned" style of deformation, in accordance with previous studies (e.g. Pérez-Estaún *et al.*, 1994). On the other hand, positive anisotropy in the hinterland is usually associated to horizontally sheared fabrics, which in the case of the Variscan orogen could either be caused during pervasive deformation associated to plate duplication in convergence (e.g. Martínez-Catalán et al., 1997), or by syn- or post-collisional localized crustal extension, as reported for the Mondoñedo thrust (Marcos, 2013). Ductile crustal flow has also been invoked as the origin of positive RA anomalies, mostly in the middle and lower crust (Lynner *et al.*, 2018), and this has also been suggested to occur at the late stages of the Varican orogeny for a significant part of the hinterland of the Variscan orogeny in NW Iberia (Llana-Fúnez and Marcos, 2007).

6 Conclusions

In this study, we used group velocity measurements from Rayleigh and Love waves to derive the first S-wave velocity model of the continental crust in NW Iberia. An independent inversion of the Rayleigh and the Love group velocity estimations was done to create a vertically polarized S-wave model and a horizontally polarized S-wave model, respectively. The first model displayed velocities between 2.3 and 3.6 km/s, while the velocities of the latter ranged between 2.6 and 4.2 km/s. Both models clearly depict the contrast in physical properties between the rocks located in the external (CZ) and the internal (WALZ, CIZ and GTOMZ) zones of the Iberian Variscan orogen, delineating the core of the large Western European Variscan belt: the core of the Ibero-Armorican arc at the Cantabrian Zone. Moreover, the high resolution of the models and the good existing geological knowledge of the area allowed us to associate the observed velocity changes to the variation of the bulk properties of the rocks caused by other first-order orogenic structures, such as the basal detachment of the CZ or the presence of a pre-Variscan basement in the NA, or sedimentary features such as the presence of syn-orogenic wedges in the CZ.

The crustal radial anisotropy 3-D model has been proved useful to identify and characterize the different sectors and units that form an orogenic belt and show a good correspondence with the dip of the main structures. The anisotropy pattern can also be associated to the deformation history of the orogen. A negative anomaly is found in the foreland fold-and-thrust belt, which can be related to the stack of thrusts and vertical fabrics caused by compression. In the internal areas, widespread positive anisotropy is accounted. The radial anisotropy magnitude oscillates between -4 and 15 % and increases with depth. Analyzing the depth pattern, we concluded that, at < 8 km depths, the alignment of Alpine fracture systems controls the anisotropy. At depths > 8 km, high and positive radial anisotropies are observed, caused by the presence of subhorizontal alignments of grains and minerals, possibly in relation with shearing in the basal detachment of the CZ and/or other shear zones formed during the development of the Variscan orogeny.

Data and resources

The datasets presented and used in this study were collected using a seismic network funded by project GEOCANTABRICA-COSTA (https://doi.org/10.7914/SN/YR_2019) and it can be released to the public on demand at GEOCANTABRICA@ftp.geol.uniovi.es. The deployment of the IberArray broadband seismic network (https://doi.org/10.7914/SN/IB) was part of the CONSOLIDER CSD2006-00041 (Geosciences in Iberia: Integrated studies on Topography and 4-D Evolution) grant from the Spanish Ministry of Science and Innovation. This dataset is available online at the ORFEUS Data Center (http://www.orfeus-eu.org/). Data processing used the codes: Corr_stack_v04.1 (Schimmel et al., 2011) and Computer Programs in Seismology v3.30 (Herrmann, 2013). Figures were drafted using Generic Mapping Tools (Wessel and Smith, 1998) and QGIS 3.10.

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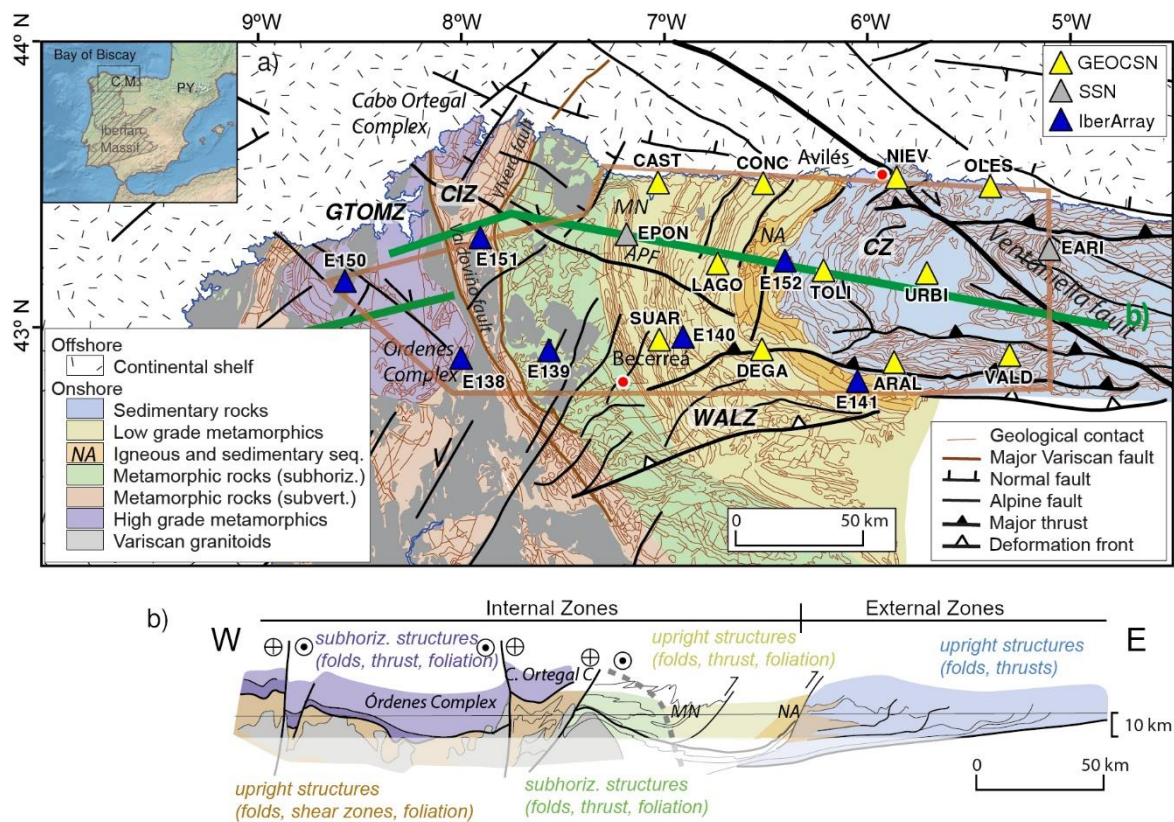
Figures

Figure 1: (a) Simplified geological map of the study area showing a lithological zonation of the Iberian Variscan orogen in NW Spain. The green line marks the position of the cross-sections in Figs. 1b, 8 and 9, while the brown polygon represents the limits of the maps in Figs. 4, 6 and 7. Triangles depict the location of the seismic stations that were used in this study. Their colours indicate the seismic network to which they belong. (Inset) Location of the investigated area within the Iberian Peninsula and the Variscan Iberian Massif. CZ: Cantabrian Zone, WALZ: West-Asturian Leonese Zone, CIZ: Central-Iberian Zone, GTOMZ: Galicia-Trás-os-Montes Zone. APF: As Pontes fault, MN: Mondoñedo nappe, NA: Narcea antiform, VF: Ventaniella fault, PY: Pyrenees, CM: Cantabrian Mountains. Background geology onshore is taken from the continuous geological map of Spain at 1:1000000 (IGME). (b) General cross-section across the northwestern Iberian Variscan belt (based on Pérez-Estaún *et al.*, 1991). The opaque part of the profile highlights the sampled depth in this study.

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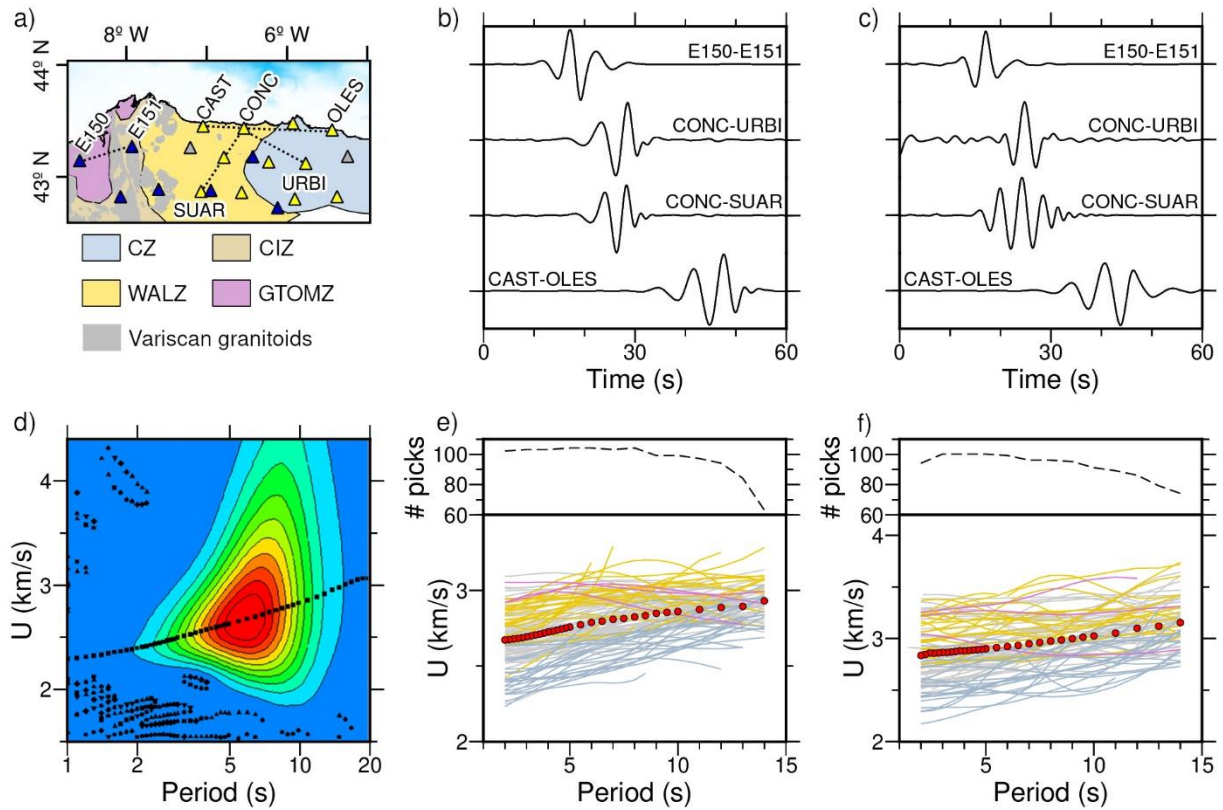
Figure 2: (a) Location of the displayed interstation paths over a simplified map of geological domains. Triangles represent the position of the seismic stations using colour codes in Fig. 1. **(b)** Examples of stacked vertical and **(c)** transverse component symmetric EGFs. Rayleigh and Love waves are the dominant phases, respectively. **(d)** Example of MFA surface for the CONC-URBI (vertical component) station pair, showing the fundamental mode Rayleigh-wave group velocity dispersion curve. **(e)** Rayleigh- and **(f)** Love-wave group velocity dispersion curves showing the number of velocity estimations used in the inversion as a function of the period. Dispersion curves obtained for paths fully contained within the main geological domains are highlighted using the colours in Fig. 2a, except for paths crossing the CIZ, GTOMZ or both, which are indistinctly marked in pink. Grey dispersion curves represent mixed paths that span through more than one region. Red circles illustrate the average dispersion values used in the computation of the initial regional S-wave velocity models. U: group velocity.

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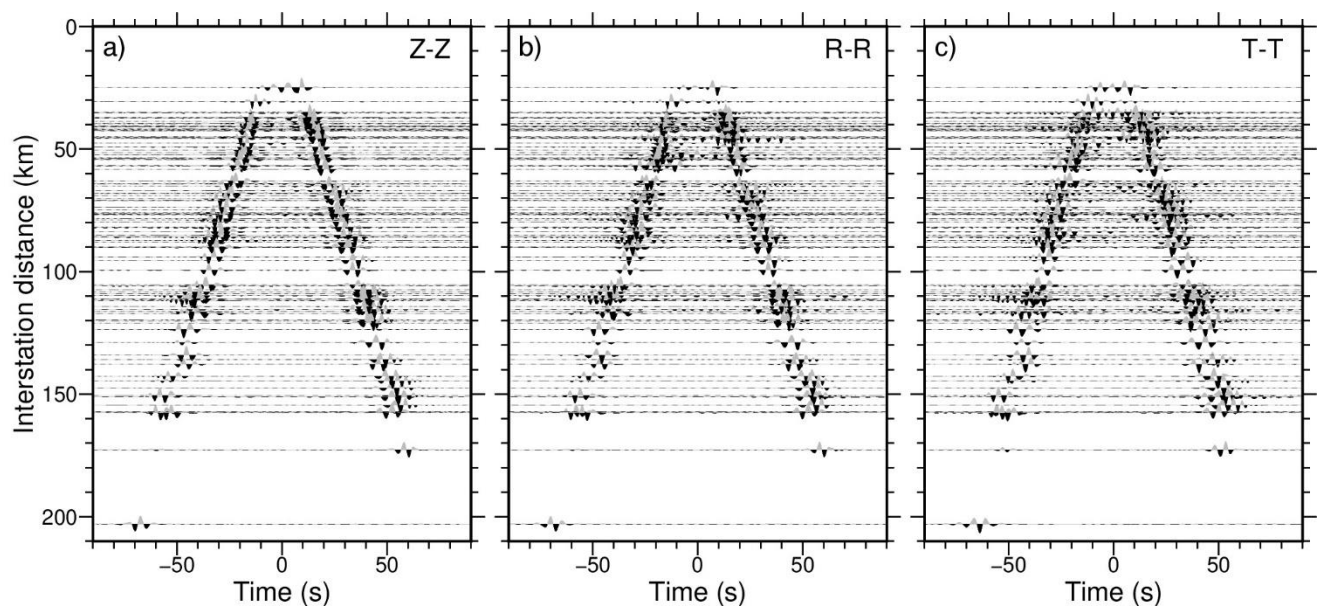
Figure 3: Record sections showing the EGFs of all the available station pairs obtained from the cross-correlation of the a) vertical (Z-Z), b) radial (R-R) and c) transverse (T-T) components.

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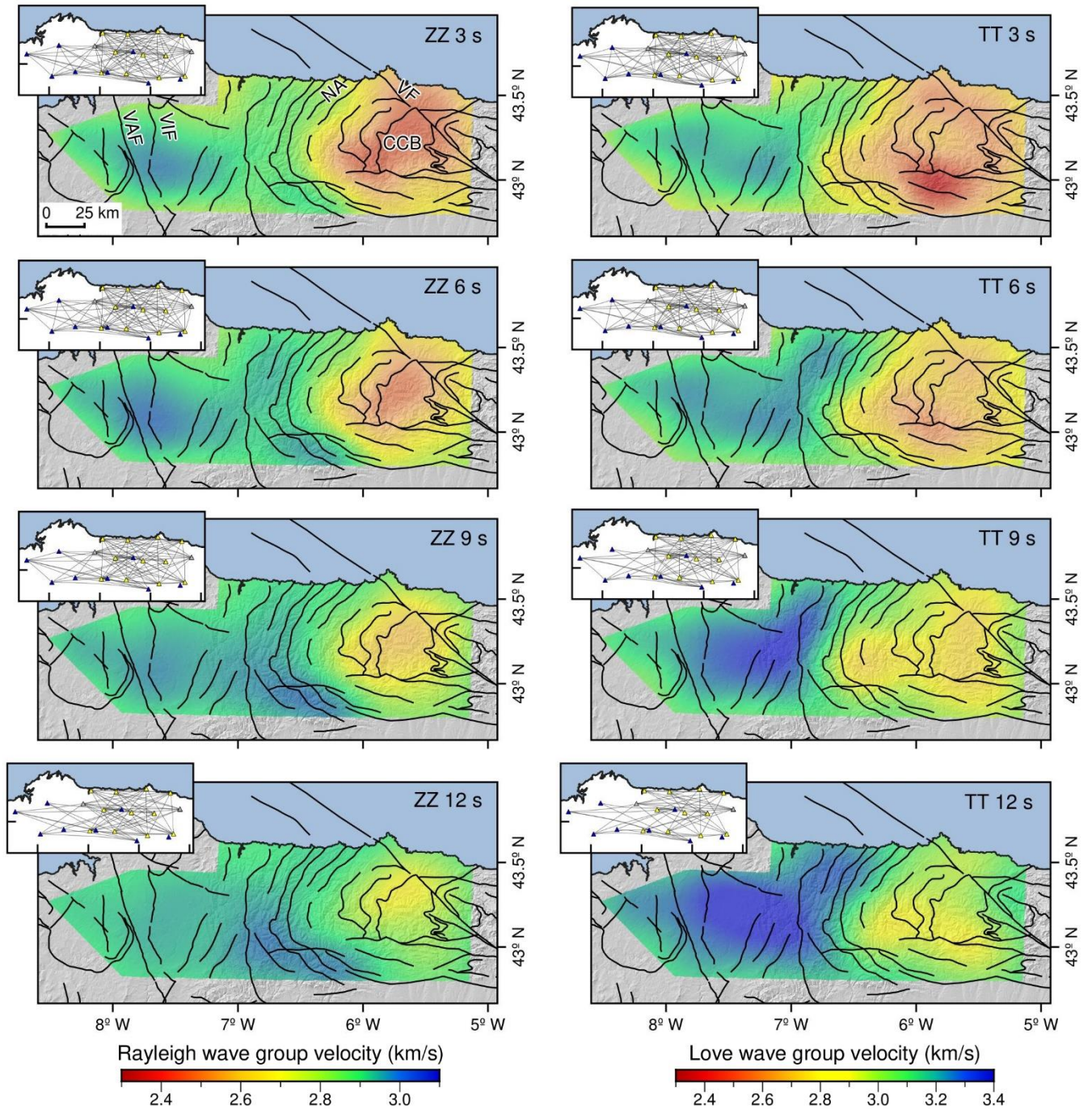
Figure 4: Rayleigh (left panels) and Love (right panels) group velocity tomographic maps for 3, 6, 9 and 12 s periods. Black lines represent the main Alpine and Variscan faults. Insets depict the interstation paths used in each inversion. CCB: Central Coal basin, NA: Narcea antiform, VF: Ventaniella fault, VIF: Vivero fault, VAF: Valdoviño fault.

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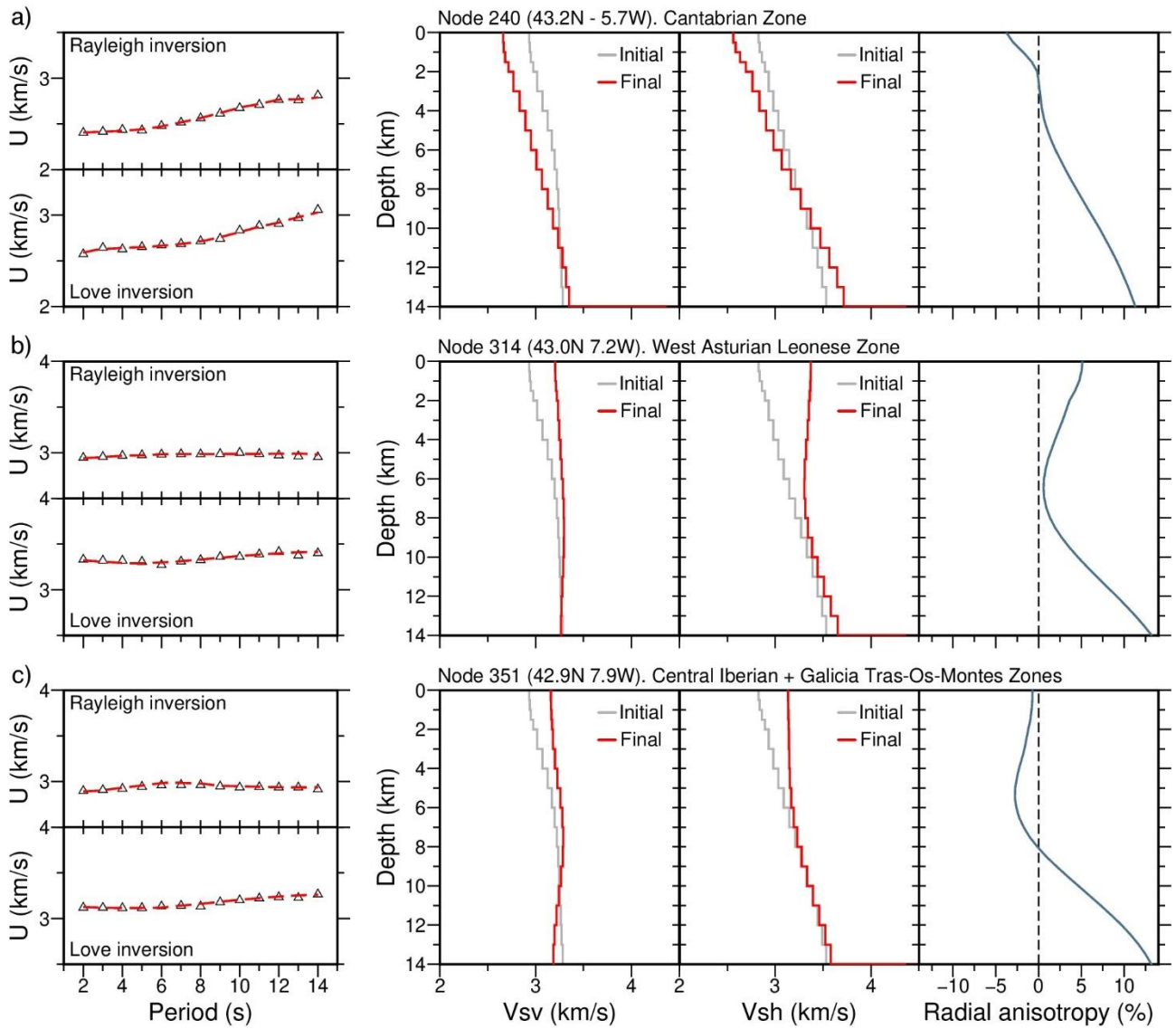
Figure 5: Depth inversion results in three nodes of the grid located in different geological domains of the Iberian Variscan orogen. Left panels display the observed group velocity data (triangles) and the inverted model fit to the dispersion (dashed red line). Central panels show the initial and final S-wave velocity models from Rayleigh (left) and Love (right) waves, respectively. Panels to the right depict the radial anisotropy magnitude calculated with Equation 1. (a) Example of inversion result in the Cantabrian
875 **Zone. (b) Example of inversion result in the West Asturian Leonese Zone. (c) Example of inversion result in the Central Iberian / Galicia Tras-Ós-Montes Zones. U: group velocity.**
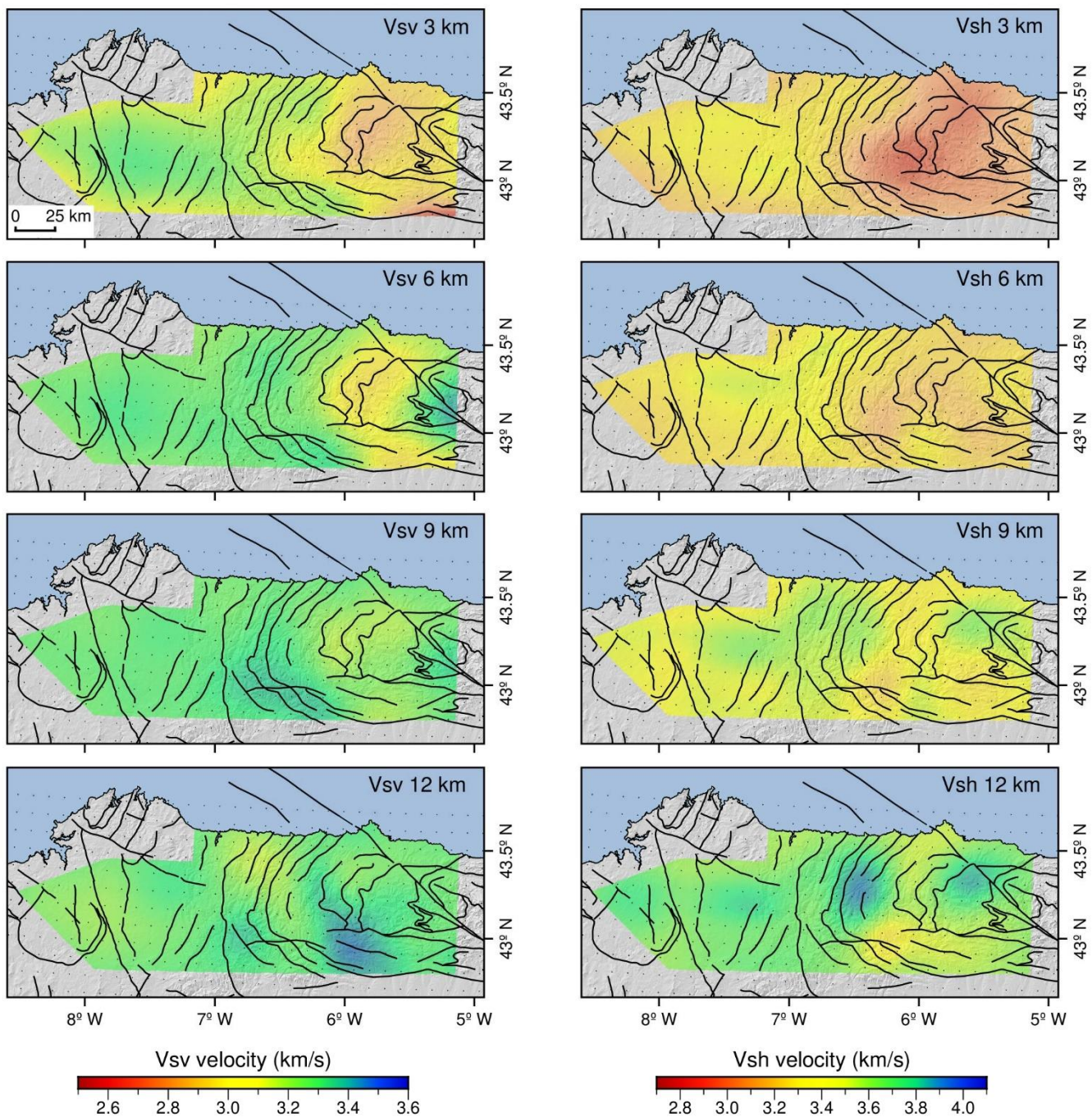
Figure 6: Inverted V_{sv} (left panels) and V_{sh} (right panels) velocity tomographic maps for depths of 3, 6, 9 and 12 km. Black lines represent the main Alpine and Variscan faults. The point grid marks the position of each inverted Vs depth model.
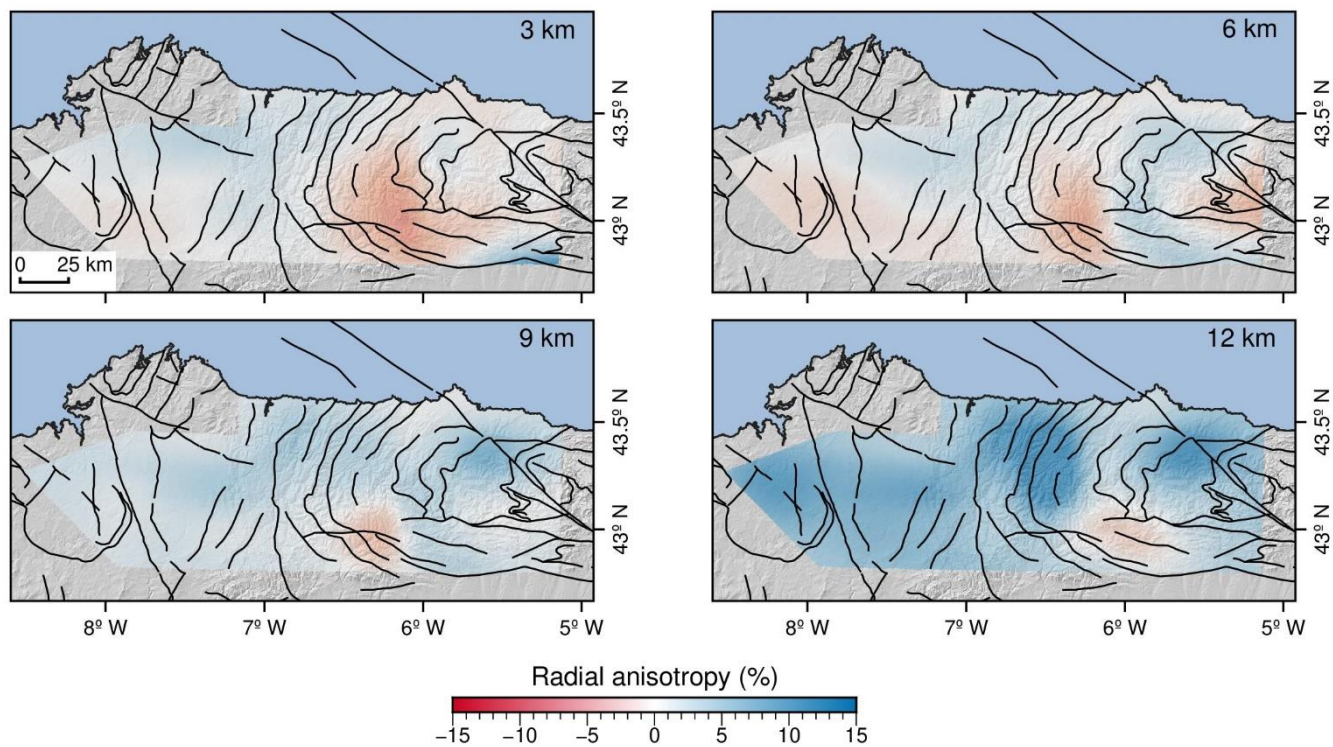
Figure 7: Radial anisotropy maps for depths of 3, 6, 9 and 12 km. Black lines represent the main Alpine and Variscan faults.

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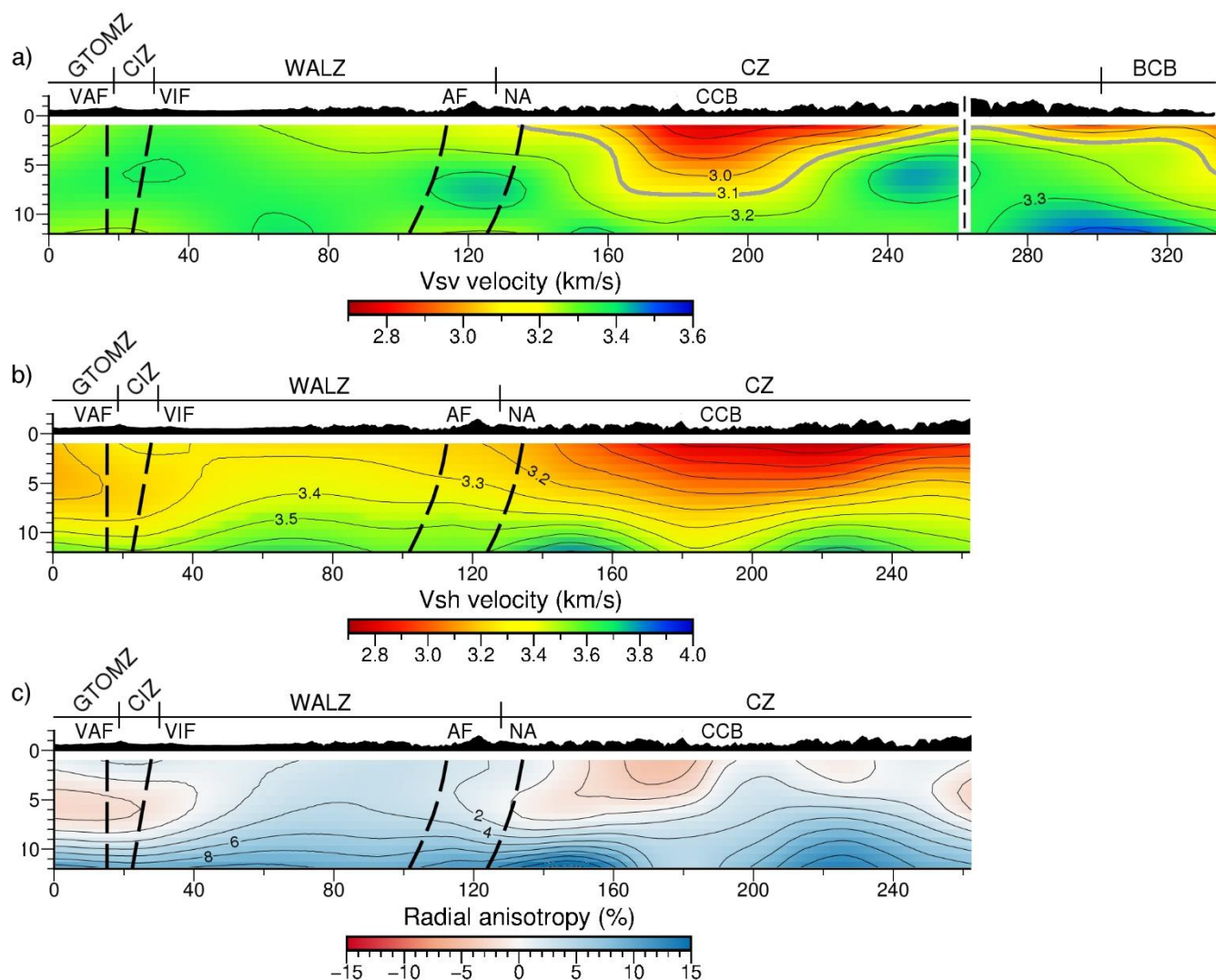
Figure 8: Cross-sections through the (a) V$_{sv}$, (b) V$_{sh}$ and (c) radial anisotropy models (location in Fig. 1), showing the topographic relief and the trace at depth of the structures that constitute the limits of the main geological domains. The V$_{sv}$ cross-section has been extended to the East using the model in Acevedo et al. (2019). The 3.1 km/s V$_{sv}$ velocity contour, which marks the depth of the cover – pre-Variscan basement boundary, is highlighted in grey. CCB: Central Coal basin, NA: Narcea antiform, AF: Allande
890 **fault, VIF: Vivero fault, VAF: Valdoviño fault.**
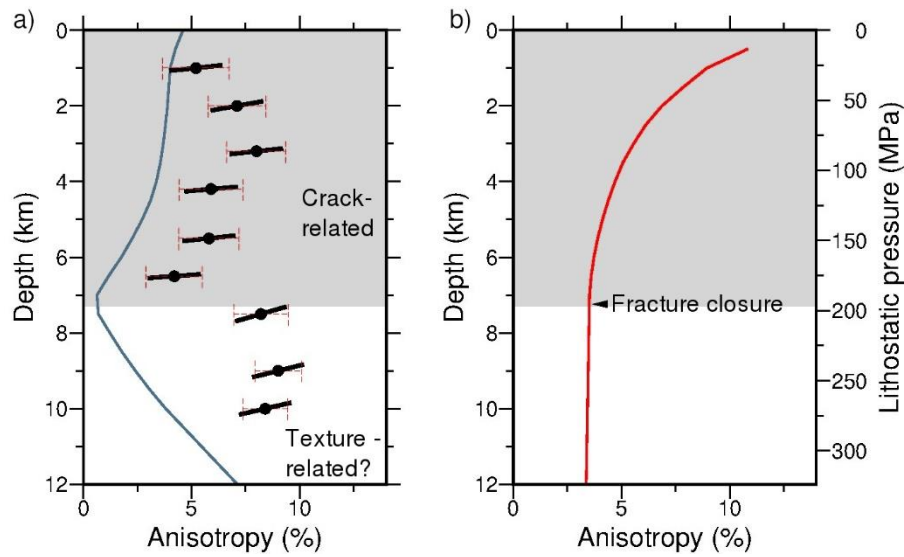
Figure 9: (a) Comparison between the absolute value of the radial anisotropy magnitude measured in a representative node of the CZ (blue continuous line) and the azimuthal anisotropy estimations (black dots) from Acevedo et al. (2020). The orientation of the black bars depicts the azimuthal fast directions. Error bars (red dashed lines) represent two standard deviations of the anisotropy magnitude. The period-depth conversion of the azimuthal anisotropy determinations is based on the sensitivity kernels in Acevedo *et al.* (2019). **(b)** General dependence of seismic anisotropy in rocks with respect to lithostatic pressure (based on laboratory measurements of S-wave velocities obtained in high-grade metamorphic rocks from NW Spain in Brown *et al.*, 2009). The grey area indicates in both panels the depths at which seismic anisotropy is controlled by the existence of open cracks.

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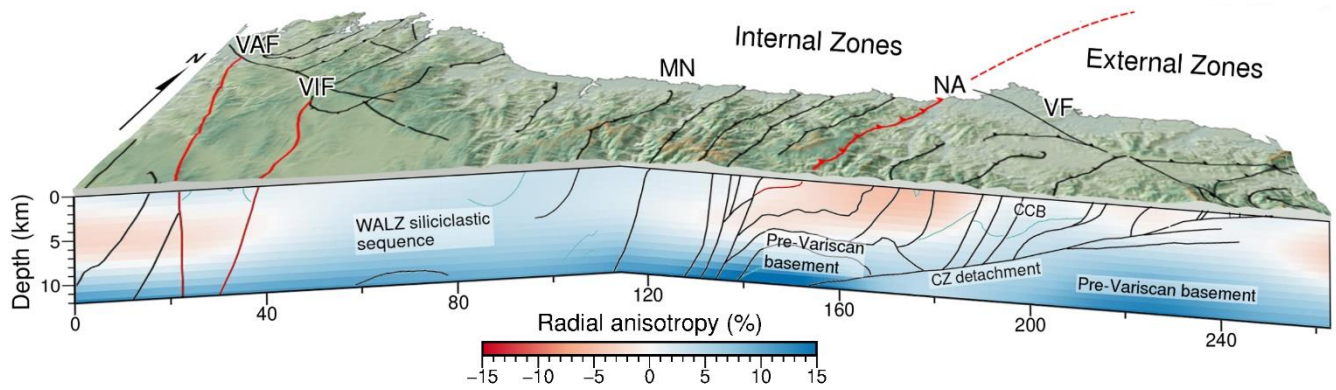
Figure 10: Schematic 3D block showing the upper-crustal radial anisotropy variation across an E-W transect of the upper crust in NW Iberia. The main structural features are superimposed, based on Pérez-Estaún et al. (1991). Red lines indicate the structures that bound the main Variscan domains displayed in Fig. 1. The location of the cross-section is highlighted in Fig. 1a. CCB: Central Coal basin, VF: Ventaniella fault, MN: Mondoñedo nappe, VIF: Vivero fault, VAF: Valdoviño fault.