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1 **The mantle transition zone dynamics as revealed through** 2 **seismic anisotropy.**

3 J.-P. Montagner¹, G. Burgos³, Y. Capdeville², E. Beucler², A. Mocquet²

¹ *Seismology Laboratory, Institut de Physique du Globe UMR-CNRS 7154, 1 rue Jussieu 75238 Paris Cedex 05, France*

² *Laboratoire de Planétologie et Géodynamique, Université de Nantes, France*

³ *C.E.A., DAM, DIF, 91297 Arpajon Cedex, France*

4 5 **SUMMARY**

6 The mantle transition zone (MTZ) of the Earth lies between 410 and ~1000 km in depth
7 and has a key role in mantle convection processes. In particular, the discontinuity at 660
8 km and its associated endothermic mineralogical transformation can slow or inhibit the
9 passage of matter between the upper and lower mantle. The MTZ thus acts as a boundary
10 layer within the mantle. The depth variations of radial and azimuthal seismic anisotropies
11 enable the detection of boundary layers within the mantle. However, the 3D imaging
12 is difficult due to the lack of sensitivity of surface waves of fundamental modes, and
13 the poor global coverage of this depth range by body-wave data. We present a new 3D
14 general anisotropy model (both radial and azimuthal anisotropies) of the mantle down to
15 1200 km in depth using surface-wave overtone datasets. We find that there is little seismic
16 anisotropy in most of the MTZ, except below subduction zones around the Pacific Ocean
17 and, more surprisingly, in a large area beneath eastern Eurasia where the Pacific subduct-
18 ing plate is stagnant. Seismic anisotropy is usually associated with intense deformation
19 processes but also possibly to water transportation or to fine layering. This significant
20 anisotropy in this part of MTZ might reveal a large water 'reservoir' associated with hy-
21 drous minerals or a strong stratification. It reflects a complex history beneath central Asia,

22 where the Tethys, Izanagi and Pacific plates appear to have strongly interacted during the
23 last 100 My, having subducted in orthogonal directions under the Asian continent, with
24 the Tethys plate descending into the lower mantle, and the Izanagi plate remaining stag-
25 nant in the MTZ. The Asian continent is the only region in the world where subducting
26 slabs originating from different plates can interact. This unique slab distribution might
27 explain why some plates descend while others remain in the lower transition zone.

28 **Key words:** Anisotropy, Surface waves, Tomography, Mantle Transition Zone

29 **1 INTRODUCTION: TRANSITION ZONE STRUCTURE AND ANISOTROPY**

30 The role of the mantle transition zone (MTZ) in the global dynamics of the Earth has been debated
31 for many decades, and is related to the long-standing debate on a whole mantle versus two-layer
32 convection. Here, we consider the MTZ in a broad sense, as located around the 660-km discontinuity,
33 and divided into the upper transition zone (410-660 km) UMTZ and the lower transition zone (660-
34 1000 km) LMTZ. There is no doubt thanks to seismic tomography, that plates can descend through the
35 MTZ into the lower mantle in most subduction zones, although their final fate can be diverse (van der
36 Hilst et al. 1997; Fukao et al. 2002). The transfer budget between material subducting into the lower
37 mantle and material remaining in the upper mantle is poorly known, which makes the nature of the
38 MTZ controversial. There were also seismic observations showing a change in seismic properties
39 around 920km depth since Kawakatsu & Niu (1994). A change in the shear velocity spectrum was
40 recently found around 1000km depth (Durand et al. 2017), which might correspond to the lower limit
41 of LMTZ. Are the 410 km and 660 km discontinuities only due to phase changes and mineralogical
42 transformation, or is the MTZ a chemical boundary between the upper and lower mantles that strongly
43 limits the flow between them?

44 An efficient way to investigate the mantle flow circulation is to map the seismic anisotropy for this
45 depth range (Montagner & Kennett 1996; Karato 1998). Seismic anisotropy is related to the depen-
46 dence of seismic wave velocity on the direction of propagation. It can have multiple causes at micro-
47 scopic and macroscopic scales. Most mantle minerals are strongly anisotropic (Mainprice 2015), and
48 anisotropy is investigated with respect to the symmetry axes of minerals. Following the pioneering
49 work of Nicolas et al. (1973, 1980), the fast axis of olivine, the main constituent of the uppermost
50 mantle, tends to align in the flow direction (Christensen & Lundquist 1982), according to its lattice or
51 crystal preferred orientation (later on referred to as LPO or CPO). This microscopic anisotropic prop-
52 erty can remain at very large scales, provided that the deformation field is coherent at scales larger

53 than the seismic wavelength, opening the possibility to map mantle convection through the orientation
 54 of seismic anisotropy. In the MTZ, olivine and pyroxenes transform into wadsleyite (410-520km depth
 55 range), ringwoodite (520-660km depth range) and garnet, and ultimately into Mg and Ca perovskite
 56 (Mg-perovskite now named Bridgmanite), ilmenite and ferropicicase (also named magnesiowüstite).
 57 The mechanisms of alignment producing this intrinsic anisotropy in the MTZ are still being debated
 58 (Tommasi et al. 2004; Kawazoe et al. 2013). Other processes can give rise to seismic anisotropy at
 59 these great depths (named extrinsic anisotropy), such as fluid inclusion oriented distribution (although
 60 this might be a secondary effect in the deep mantle), or fine layering of isotropic material with very
 61 different velocities, which also creates a specific case of anisotropy: radial anisotropy (Backus 1962).
 62 Radial anisotropy is characterized by five independent elastic parameters (Love 1944) and no horizon-
 63 tal azimuthal velocity variations. The corresponding medium is a transversely isotropic medium with
 64 a vertical symmetry axis, which is usually known as the vertical transverse isotropy (VTI) medium.
 65 An example of such a finely layered or laminated medium with partial melting is illustrated by the
 66 millefeuilles model of Kawakatsu et al. (2009) of the asthenosphere. Even though the interpretation
 67 of observed anisotropy is non-unique, for large-scale mapping of fast directions of velocities, we need
 68 both radial and azimuthal anisotropies, which provide unique information on convective flows in the
 69 mantle (Tanimoto & Anderson 1984; Montagner 2002) and which can be compared to geodynamic
 70 models (e.g. Gaboret et al. (2003); Becker et al. (2003); Simmons et al. (2009); Long & Becker (2010)
 71 ...).

72 Seismic data enable us to observe these two main kinds of anisotropy: radial anisotropy, as above,
 73 which only requires a VTI medium to explain the Rayleigh-Love discrepancy (Anderson 1961), and
 74 azimuthal anisotropy of different body waves since the first observations on Pn (Hess 1964) and SKS
 75 splitting (Vinnik et al. 1984; Silver & Chan 1991), and surface waves at the regional (Forsyth 1975)
 76 and global (Montagner & Tanimoto 1991) scales. A VTI model is characterized by 5 parameters
 77 A, C, F, L, N (Love 1944), but for the inversion technique, they are recombined as $V_{PH}, V_{SV}, \phi =$
 78 $C/A = (V_{PV}/V_{PH})^2, \xi = N/L = (V_{SH}/V_{SV})^2, \eta = F/(A - 2L)$. The corresponding sensitivity
 79 kernels are modified accordingly. Among the five independent elastic parameters of a VTI medium,
 80 the most easily accessible parameters are the V_{SV} velocity and the ξ parameter, which expresses the
 81 difference in the velocities between horizontally and vertically polarized S-waves ($\xi = (V_{SH}/V_{SV})^2$). A
 82 correct account of the crustal structure is mandatory to avoid a bias between the shallow structure
 83 and the deep structure, especially the radial anisotropy. This non-linear effect was first made evident
 84 by Montagner & Jobert (1988) and later on carefully investigated by Marone & Romanowicz (2007);
 85 Ferreira et al. (2010); Chang et al. (2015); Chang & Ferreira (2017). We make use of the crustal model
 86 of Burgos et al. (2014) derived from the joint inversion of short period group and phase velocities

87 (periods of 15-40s). This kind of simple ξ anisotropy was introduced for the 1D spherically symmetric
 88 preliminary reference earth model (PREM) in the uppermost 220 km of the mantle (Dziewonski &
 89 Anderson 1981). Azimuthal anisotropy cannot be explained by a VTI model and requires instead a
 90 more complex model of anisotropy. When derived from surface waves, additional eight parameters are
 91 necessary (Montagner & Nataf 1986), with the best-resolved azimuthal parameters being the modulus
 92 G and its associated angle Ψ_G related to the azimuthal variation of V_{SV} .

93 For several decades, there has been good observational evidence for radial and azimuthal seismic
 94 anisotropies in the top 300 km of the upper mantle and in the bottom D''-layer. In the transition zone,
 95 1D radial anisotropy was introduced to explain eigenfrequency data (Montagner & Kennett 1996),
 96 and it was extensively investigated later (Beghein et al. 2008). Since then, several global tomographic
 97 models (Kustowski et al. (2008); Panning et al. (2010); Moulík & Ekström (2014); Auer et al. (2014);
 98 French & Romanowicz (2014); Chang et al. (2015)) have displayed the radial anisotropy parameter ξ
 99 and have shown that there might be some secondary maximum (although small) of radial anisotropy in
 100 the MTZ. Some hints of azimuthal anisotropy in the MTZ (Vinnik & Montagner 1996; Wookey et al.
 101 2002) have been provided by different kinds of body-wave data (e.g., P-to-S receiver functions, shear-
 102 wave splitting), which have demonstrated lateral variations in anisotropy in the transition zone. On
 103 a global scale, the first long-wavelength azimuthal anisotropic 3D structure in the MTZ (Trampert &
 104 van Heijst 2002) has been obtained by the inversion of Love wave overtone data, which have a limited,
 105 although significant, radial resolution in this depth range. Its lateral resolution, more than 5000 km,
 106 was too poor to gain insights into the coupling and transfer of matter between the upper and lower
 107 mantles. More recent tomographic models displayed azimuthal anisotropy distributions down to the
 108 transition zone (Yuan & Beghein 2013; Debayle et al. 2016; Schaeffer et al. 2016), but the agreement
 109 between them is rather poor (Huang et al. 2019). The discrepancies might be due to the fact that rms
 110 amplitude of the lateral variations in MTZ was found to be very small (about 1%, within the error
 111 bars), and to be much smaller than in the uppermost and lowermost mantle (the D''-layer).

112 **2 ANISOTROPY TOMOGRAPHY FROM OVERTONE DATA**

113 In this study, we combine two datasets of surface wave (Rayleigh and Love waves) overtones (higher
 114 modes), (Beucler & Montagner 2006; Visser et al. 2008) using different methods and different data.
 115 The merged dataset improves the global coverage of the Earth, and enables us to derive a global 3D
 116 map of the radial and azimuthal seismic anisotropies in the MTZ down to the mid-mantle (~ 1200
 117 km). The merged dataset improves the global coverage of the Earth. It enables us to derive a global
 118 3D map of the radial and azimuthal seismic anisotropies in the MTZ down to the mid-mantle (~ 1200
 119 km). This model has a lateral resolution of 1000 km down to a depth of 1200 km. It also includes phase

120 velocities of the higher modes of Rayleigh waves up to order 6, and higher modes of Love waves up to
 121 order 5. In the period range of 35 s to 250 s, the number of measurements of the merged dataset varies
 122 from 70000 for the fundamental Rayleigh mode (and 50000 for the fundamental Love mode) down
 123 to 30000 for the 6th Rayleigh higher mode (and 8500 for the 5th Love higher mode). The measure-
 124 ment techniques are different for each of the datasets, but both include data uncertainties. Beucler &
 125 Montagner (2006) uses the roller-coaster algorithm (Beucler et al. 2003) that clusters close events and
 126 extracts the individual phase velocities for each overtone by non-linear inversion. Visser et al. (2008)
 127 uses a model-space search approach (Yoshizawa & Kennett 2002). Several tests have been performed
 128 to check the compatibility of each of the datasets, by doing separate inversions (regionalizations of
 129 phase velocities at different periods), and joint inversions of the datasets (Burgos 2013). The path
 130 and the azimuthal coverages for the fundamental modes of Rayleigh and Love waves are presented
 131 in Figures 1 and 2 of Burgos et al. (2014). Path and azimuthal coverages for overtones have similar
 132 distributions but with a smaller density.

133 We did not find large discrepancies between the two datasets, and the data of the higher mode
 134 phase velocities were jointly regionalized by taking into account their respective error bars (Figures
 135 A1 and A2). The 3D anisotropic tomographic method is an improved version of a classical two-
 136 step inversion procedure (Montagner 1986) based on gradient least-squares optimization (Tarantola
 137 & Valette 1982). The first step consists in regionalizing the measured phase (or group) velocities
 138 at different periods along the different paths in order to get the global geographical distribution of
 139 velocities. The second step is devoted to the inversion at depth of these velocity distributions in order to
 140 retrieve the 3D-distributions of seismic velocities and anisotropic parameters. The complete technique
 141 is detailed in the Appendix. The output tomographic model consists of the V_S velocity and some robust
 142 anisotropic parameters, $\xi = (\frac{V_{SH}}{V_{SV}})^2$, the S-radial anisotropy, G , with Ψ_G expressing the horizontal
 143 azimuthal dependence of V_{SV} velocity ($V_{SV} = V_{SV0} + G \cos(2(\Psi - \Psi_G))$), and to a lesser extent, E ,
 144 with Ψ_E expressing the azimuthal dependence of V_{SH} velocity ($V_{SH} = V_{SH0} + E \cos(4(\Psi - \Psi_E))$).

145 Sensitivity kernels computed in PREM show why higher modes make it possible to obtain good
 146 depth resolution down to 1200 km, well below the upper (410-660 km) and lower (660-1000 km)
 147 MTZ, which are the main focus of the present study. Figure A3 (see Appendix) shows examples of the
 148 sensitivity kernels of Rayleigh and Love waves, respectively, with respect to SV-wave and SH-wave
 149 velocities for different higher modes. The maximum sensitivity at a given period increases with the
 150 order n of the higher mode. For example, at 51 s, the 4th Rayleigh higher mode (Figure A3a) has a
 151 maximum sensitivity in V_{SV} at 800 km in depth. The sensitivity of Rayleigh waves to V_{SH} is small,
 152 whereas the higher modes of Love waves can also be very sensitive to V_{SV} . This property was used
 153 by Trampert & van Heijst (2002) to retrieve SV-wave anisotropy in the MTZ by using only the Love

154 wave data. Parameter G (and similarly, parameter E) associated with the $2 - \psi$ (and $4 - \psi$) azimuthal
 155 dependence of SV-waves (resp. SH-waves) has the same kernel as V_{SV} (resp. V_{SH}), and therefore it
 156 can also be recovered down to the mid-mantle at 1200 km in depth (even though only the results down
 157 to 800 km in depth are presented here). V_{SV} , V_{SH} , G and E are the most robust parameters that can
 158 be retrieved by inversion at depth of the higher mode phase velocity distributions. In the second step
 159 of the tomographic technique, the inversion at depth of all of the azimuthal dependence functions of
 160 the Rayleigh and Love wave local phase velocities is performed at every point, to retrieve these most
 161 robust elastic parameters. The local crustal structure was previously derived by Burgos et al. (2014).
 162 PREM (Dziewonski & Anderson 1981) is used below the crust as the starting reference model. But
 163 after inversion, the global average ξ_{ref} of ξ is the new reference radial anisotropy model.

164 The tomographic model at 100 km in depth (Figure 1 Top) shows what we expect in terms of the S-
 165 wave velocity and azimuthal anisotropy distributions. The ridges and back-arc zones are slow, the old
 166 continents are very fast. For most oceanic plates, the azimuthal anisotropy correlates well with the plate
 167 motions, and the radial anisotropy parameter ξ is close to the PREM value (we only represent $\delta\xi$, the
 168 deviation with respect to PREM). Deeper in the MTZ, the amplitude of the heterogeneities is largely
 169 decreased, but the slab signature is perfectly visible, not only in the upper transition zone UMTZ
 170 (Figure 1 Middle, left, 500-km depth), but also in the lower transition zone LMTZ (Figure 1 Bottom,
 171 left, 800-km depth). The azimuthal anisotropy parameter G is small in most of the oceanic areas (with a
 172 variance of 0.3% at 500km and 0.1% at 800km, smaller than the error bars $\sim 0.5 - 1\%$), but relatively
 173 large ($> 1\%$) in most of the subduction zones, and surprisingly below some parts of continents, in
 174 particular below Asia. Our results differ from Yuan & Beghein (2013) who found distributed azimuthal
 175 anisotropy with a larger amplitude, but in correct agreement with Schaeffer et al. (2016); Debayle et al.
 176 (2016) who found a small rms amplitude in the UMTZ. The differences between models might be due
 177 to the joint inversion of different Rayleigh and Love wave datasets with different inversion techniques.
 178 The radial anisotropic parameter ξ in UMTZ is also much smaller than in the uppermost 300 km of the
 179 mantle, and it shows a patchy pattern. The distribution of parameter ξ (Figure 1 Right) at 500 km depth
 180 is also very patchy and slightly above the errors bars only below the subduction zones, western Africa
 181 and Central Asia. The error bars are calculated as the diagonal terms of the a posteriori covariance
 182 function of parameters (Montagner 1986). At 800 km in depth (Figure 1 Bottom), the pattern of ξ
 183 is completely different since it is almost everywhere below the error bars, but displays significant
 184 negative values below western America, central Africa and central Asia.

185 The lateral heterogeneities can be expanded in terms of the spherical harmonics and their power
 186 spectrum is computed to determine how the amplitudes of the heterogeneities in the S-wave velocity
 187 (Figure 2 Left) and azimuthal anisotropy (Figure 2 Right) vary at depth. The angular order ℓ of power

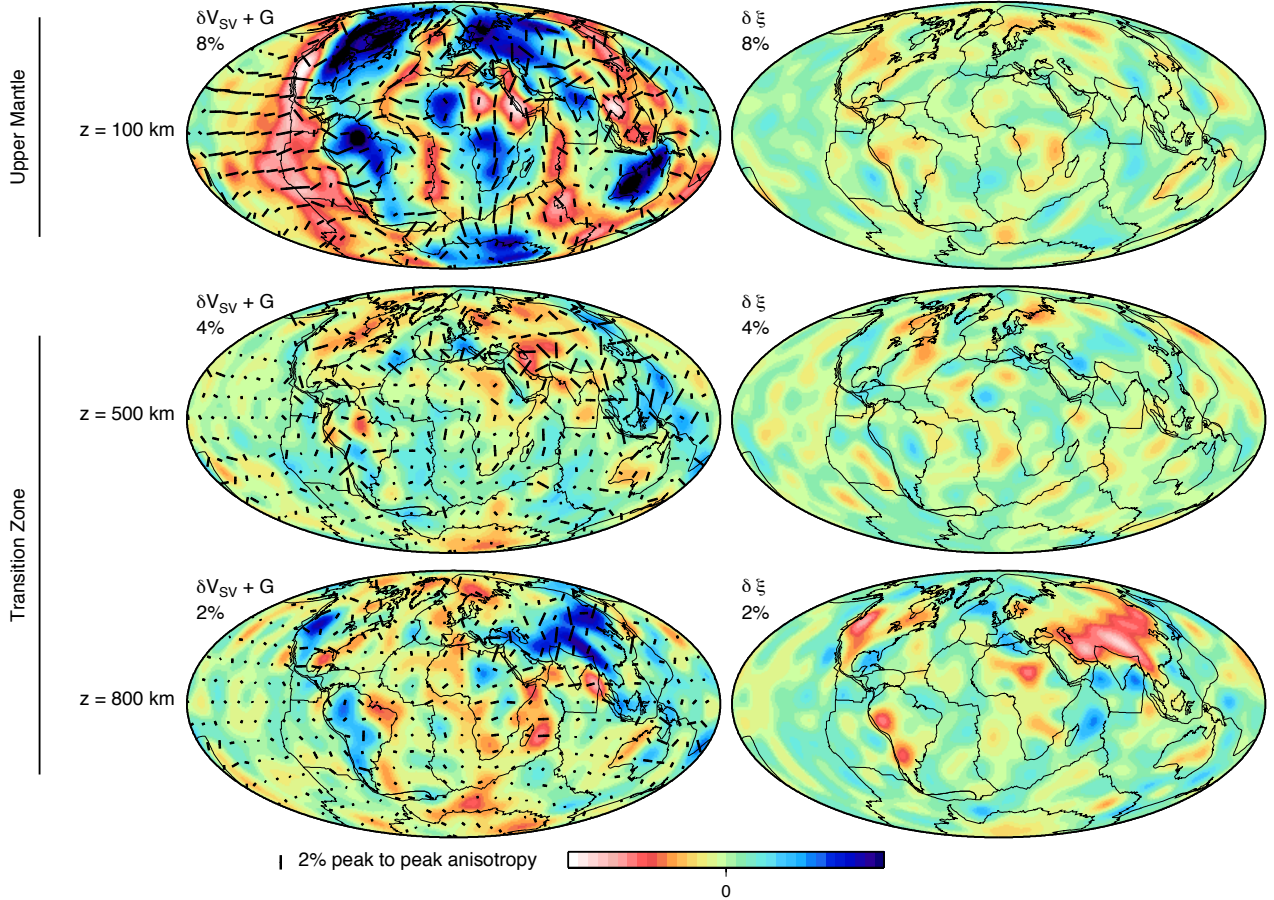


Figure 1. Tomographic images of the S-wave velocity V_{SV} , azimuthal anisotropy G (left) and radial anisotropy $\delta\xi = \xi - \xi_{ref}$ (right) in different mantle layers: The upper mantle, at 100 km in depth (top). The transition zone, at 500 km (middle) and 800 km (bottom) in depth. $\delta\xi$ is plotted with respect to an average reference model, $\xi_{ref} = 2.7\%$ at 100km, -1.5% at 500km, 0.4% at 800km. The error bars are calculated as in Burgos et al. (2014) and presented in Burgos (2013). Error bars on ξ and G are everywhere smaller than 1% and even smaller than 0.5% in well covered regions of the northern hemisphere. The color scale varies with depth and its range is defined at the top left of each figure.

188 spectra can be related to the spatial wavelength Λ through the relationship $\Lambda = \frac{2\pi a}{\ell+0.5}$, where a is
 189 the radius of the Earth. Rapid decreases in the amplitudes of the S-wave and the azimuthal parameter
 190 G are observed between 200 km and the different depths of the MTZ. Degree 2 of the spherical
 191 harmonics expansion is dominant in the UMTZ (Figure 2, 500 km, 600 km), but there is a change
 192 in the pattern of the S-wave velocity in the LMTZ at 700 km in depth, which is dominated by de-
 193 gree 1 and a decrease in the power spectrum, with the angular order ℓ related to the inverse of the
 194 spatial wavelength Λ . This is in agreement with the results of Kustowski et al. (2008), who observed
 195 a change in the heterogeneity distribution around the 660-km discontinuity. The spectral amplitude
 196 of parameter G is always smaller than the spectral amplitude of the S -wave velocity heterogeneity,

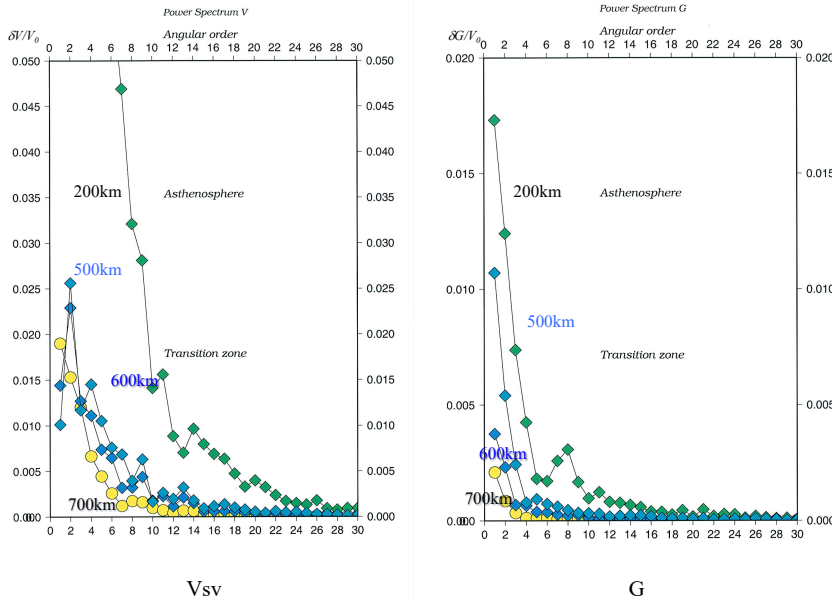


Figure 2. Power spectra of the relative variations of S_V -wave velocity (Left) and of the azimuthal anisotropy G (Right) at different depths: Asthenosphere, 200 km; Transition zone: 500 km, 600 km and 700 km, as indicated. 200km: green diamonds; 500km: light blue diamonds; 600km: dark blue diamonds; 700km: yellow circles.

197 which can be understood considering the predominant contribution of the oceanic regions away from
 198 subduction zones ($\sim 2/3$ of the Earth surface) with a small value of parameter G in MTZ. The areas
 199 where the amplitude of parameter G is significant (subducting zones and central Asia) finally have a
 200 limited influence on the global average. The distributions of azimuthal and radial anisotropies found
 201 by different tomographic models (Yuan & Beghein 2013; Schaeffer et al. 2016; Debayle et al. 2016)
 202 are in poor agreement which is due to the small amplitude of anisotropy in MTZ often within their
 203 error bars. Another approach for studying MTZ azimuthal anisotropy was proposed by Huang et al.
 204 (2019) by using long-period SS precursors. It confirms the weakness of anisotropy in most oceanic
 205 areas except around subduction zones. Consequently, we now focus on the only areas where azimuthal
 206 anisotropy is significant, above the error bars, i.e. subduction zones and Central Asia.

207 3 THE UNIQUE ANISOTROPY BENEATH CENTRAL ASIA

208 The azimuthal anisotropy in the MTZ as displayed in Figure 1 is well correlated with subducting
 209 plate slabs and is relatively small in other regions, with the noticeable exception of central Asia.
 210 Fast V_{SV} and negative ξ radial anisotropy is also observed in Canadian and Andean cordillera, but
 211 their lateral extension is much smaller than in Central Asia. The significant azimuthal anisotropy is
 212 limited to eastern Asia at 500 km in depth, but it extends through the whole of central Asia at 800
 213 km in depth (Figure 3). Unfortunately, this region is not well covered by SS precursors as investigated
 214 by Huang et al. (2019), making a comparison impossible. The puzzling azimuthal anisotropy in this

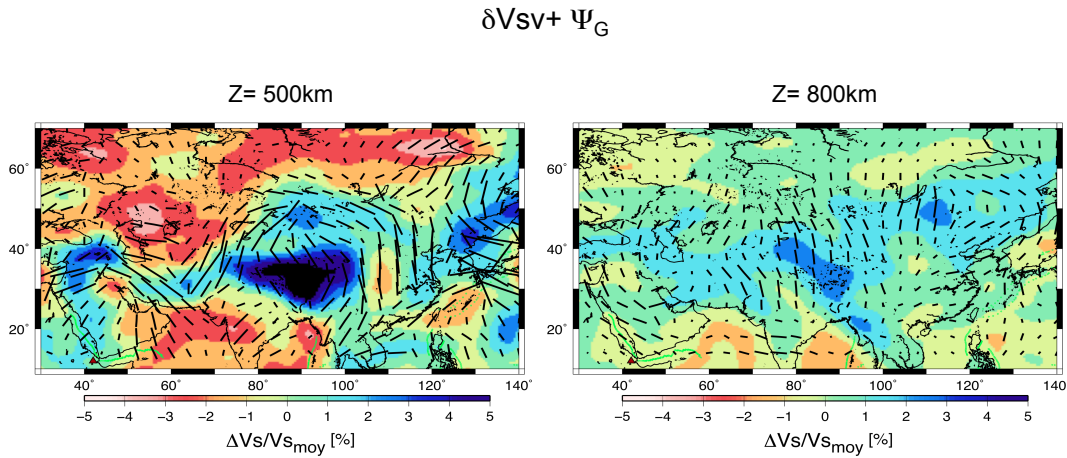


Figure 3. Tomographic images of S-wave velocity V_{SV} , azimuthal anisotropy G , in the central Asia transition zone at depths of 500 km (left) and 800 km (right).

215 region can be interpreted in terms of the mineralogy and/or the large-scale remnant plate structures
 216 and mantle convective flow. The mineralogical interpretation of the seismic anisotropy in the upper
 217 and lower MTZ is still subject of large uncertainties (e.g. the review of Mainprice (2015)). In the
 218 UMTZ (410-660 km in depth), the 410-km discontinuity, attributed to the transformation from olivine
 219 to wadsleyite (orthorhombic), might result in a decrease in the anisotropy. A weaker discontinuity at
 220 520 km in depth, e.g., Shearer (1996), has been attributed to the transformation from wadsleyite to
 221 ringwoodite. Wadsleyite (anhydrous or hydrous) can be strongly anisotropic whereas ringwoodite has
 222 a low anisotropy. This anisotropy might also be due to petrological layering caused by garnet-rich and
 223 ringwoodite-rich layers of transformed subducted oceanic crustal material (Karato 1998). However,
 224 notwithstanding some subduction zones where a tilted layering might be present, the resulting VTI
 225 medium would not cause any splitting for vertically propagating S-waves, nor produce the azimuthal
 226 anisotropy that is observed by Trampert & van Heijst (2002); Yuan & Beghein (2013); Debayle et al.
 227 (2016); Schaeffer et al. (2016) and in the present study. From 660-km to 900-1000-km deep (the
 228 LMTZ), the observed anisotropy might be due to the Bridgmanite lattice preferred orientation (and
 229 possibly to ferropericlase) that is caused by the deformation in the convective boundary layer at the top
 230 of the lower mantle (Karato 1998). Its lower limit might be related to a weak seismic discontinuity at
 231 920 km, as identified by Kawakatsu & Niu (1994) in several subduction zones. The mineralogy of the
 232 MTZ (not only subducted plates but as well other regions) is relatively complex, and different minerals
 233 were identified (Vacher et al. 1998; Anderson 2007) which can give rise to seismic anisotropy. The
 234 mineral stishovite SiO_2 that is probably present in subducting oceanic plates is strongly anisotropic,
 235 and it might contribute to the observed anisotropy. Akimotoite, which is the ilmenite form of MgSiO_3 ,

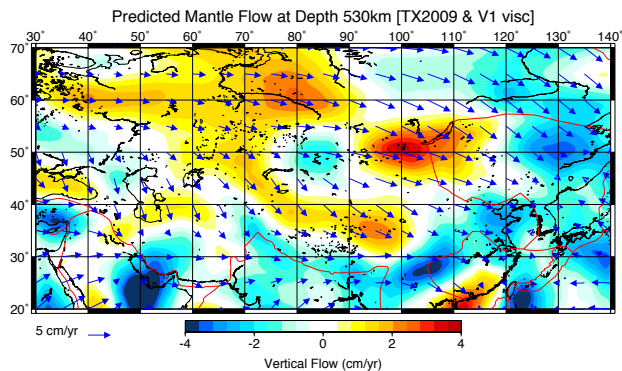
236 might dominate in cold parts of the transition zone, i.e. in areas near subduction zones (Shiraishi et al.
237 2008; Cottaar & Deuss 2016). Akimotoite has higher velocities than garnet at similar depths in hotter
238 regions of the transition zone, and can produce a local high velocity that can be misinterpreted as
239 a pile-up of subducted materials. So significant azimuthal anisotropy might be observed where the
240 akimotoite content is high. A typical structure of perovskite may be cubic, but, in the LMTZ, when
241 the crystal lattice is subject to strong distortion, silicate perovskites are in the orthorhombic crystal
242 system. Bridgmanite is anisotropic (Meade et al. 1995; Mainprice et al. 2008) and a good candidate
243 for explaining anisotropy in LMTZ since the CPO pattern and dominant slip system were carefully
244 investigated (Tsujino et al. 2016) but still controversial (Kraych et al. 2016). Ferropericlasite is nearly
245 isotropic in the LMTZ but its anisotropy strongly increases with pressure and might be as large as 40%
246 in the lower mantle.

247 A role for water can also be considered here. It was suggested that the MTZ might be a water
248 reservoir (van der Meijde et al. 2003; Bercovici & Karato 2003) that collects the water from the sub-
249 ducting plates through hydrous minerals. The effect of water on CPO and slip systems was extensively
250 investigated since hydrous wadsleyite might store a large amount of water transported by subducting
251 slabs. But there is still some debates on the dominant slip systems able to generate seismic anisotropy
252 (Tommasi et al. 2004; Kawazoe et al. 2013). A significant water content near subducted slabs (Chang
253 & Ferreira 2019) can change the rheology of minerals (i.e., their viscosity and melting (Ohuchi et al.
254 2014; Ritterbex et al. 2015, 2020)). However, Ohuchi et al. (2014) argued that, due to changes in slip
255 systems, anisotropy becomes weak if the UMTZ is hydrous and strong anisotropy near subducted slabs
256 means a dry UMTZ (Ferreira et al. 2019). So, water transportation alone cannot explain this strong
257 anisotropy and other mechanisms must be found (Ritterbex et al. 2015, 2020).

258 Nevertheless, we observe significant anisotropy in and around subducting plates, because they are
259 specific places where the physical conditions of strong deformation are fulfilled for the development of
260 the preferred orientation of minerals or of fine layering. In addition, central Asia is the continent below
261 which many oceanic plates subduct or have subducted. In particular, the Tethys plate and the Izanagi
262 plate (now extinct in the Pacific Ocean; see Lithgow-Bertelloni & Richards (1998); Scotese & Golonka
263 (1997)) subducted beneath Eurasia for more than 100 Ma (see Torsvik et al. (2010) for kinematic
264 plate reconstruction). As the Tethys and Izanagi and nowadays Pacific plates subduct in orthogonal
265 directions under the Asian continent, some strong interaction and even collision might occur between
266 these plates beneath central Asia. The Tethys plate moved northward before descending into the lower
267 mantle (van der Hilst et al. 1997), whereas the Izanagi plate became stagnant in the MTZ (Fukao et al.
268 1992).

269 In order to investigate the convective flow below Asia, we also made some comparisons with geo-

a)



b)

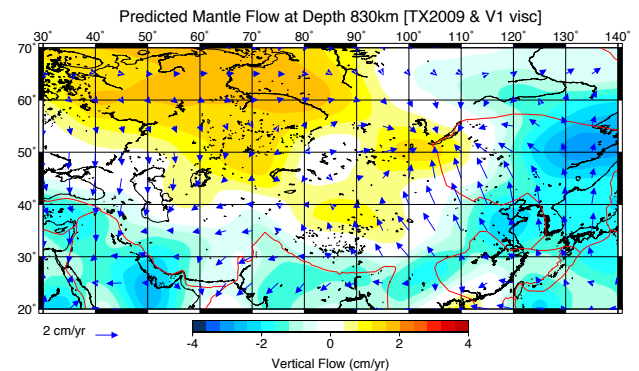


Figure 4. Flow models computed according to Simmons et al. (2009), for the depths of 500-550 km (a) and 800-850 km (b) (Courtesy of Alessandro Forte).

270 dynamic models. In the mantle circulation models of Simmons et al. (2009), it is observed (Figure 4a,
 271 b) that the flow patterns drastically change between 500 km and 800 km in depth. A more quantitative
 272 comparison of slab volume and past kinematics as done by Hafkenscheid et al. (2006) is a future path
 273 of research but is beyond the scope of this paper. The approach of Simmons et al. (2009) is based
 274 on the joint interpretation of seismic and geodynamic data (Nuvel-1 plate velocity model (DeMets
 275 et al. 2010), free air gravity data, dynamic topography) requiring mineral physical parameters link-
 276 ing seismic velocity to density perturbations in the Earth's mantle. The viscosity profile derived from
 277 convection models and GIA (global isostatic adjustment) data presents a minimum in the depth range
 278 520-650km depth and a maximum around 1500- 2000km depth. Some hypotheses used in Simmons
 279 et al. (2009) are questionable, but these simulations produce flow patterns consistent with the different
 280 patterns in azimuthal anisotropy between 500km depth and 800km depth predicted by our tomographic
 281 inversions. Beneath central Asia, the flow at 530 km in depth is primarily horizontal West-to-southeast,
 282 whereas it is primarily oriented East-to-northwest at 830 km in depth. This might correspond to the
 283 remnant of the Izanagi plate which subducted beneath east Asia until 100 Ma ago (i.e., Torsvik et al.
 284 (2010)). The directions of the anisotropy (Figure 3) and flow (Figure 4) are poorly correlated at both
 285 of these depths, with large angle differences between them. The important observation here is that
 286 the large change in anisotropy direction between the UMTZ and LMTZ is also associated with the
 287 dramatic change in the flow orientation. Further geodynamic modeling is necessary, as this depends
 288 on the chosen tomographic model and the viscosity profile. Some alternative modeling has been tried
 289 (e.g., Becker et al. (2008)), but as long as the orientation mechanisms of the minerals in the MTZ are
 290 not so well understood, there is little rationale to go further in the interpretation. In the absence of

291 consensus for the dominant mechanisms of alignment of bridgmanite-ferropericlase, it is difficult to
292 reach strong conclusions, and a better understanding of the mineralogical processes in play in the seis-
293 mic anisotropy of the MTZ will be necessary to reconcile the flow pattern directions and the observed
294 anisotropy directions.

295 So far, past ridge processes were investigated by seismic anisotropy in ophiolites (Nicolas et al.
296 1980; Nicolas & Christensen 1987; Peselnick & Nicolas 1978). The recent investigation of seismic
297 anisotropy beneath and around subduction zones (Nowacki et al. 2015; Ferreira et al. 2019; Sturgeon
298 et al. 2019) provide new insights into our understanding of convective processes and the coupling
299 between the upper and the lower mantles. What is observed at the global scale, e.g. the interaction
300 between the remnant slabs beneath the continent for central Asia is almost unique in the world, and
301 leaves an imprint through seismic anisotropy. The cluttering of the plate slabs in the MTZ might
302 explain why some of the plates descend, why some others do not, and why some are stagnant in
303 the lower transition zone. In addition to anisotropic minerals, several conditions must be fulfilled to
304 observe the seismic anisotropy in the MTZ, the large-scale deformation field, and/or the probability of
305 the presence of water carried by the subducting plates. These conditions are only present in subduction
306 zones and beneath central Asia, where past subducting plates might have accumulated not only in the
307 upper transition zone (UMTZ: 410-660 km in depth), but also in the lower transition zone (LMTZ:
308 660-1000 km in depth). The presence of subducting material present at the same time in multiple
309 depth ranges of MTZ would explain why anisotropy might be significant only in specific regions of
310 the transition zone.

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485 **Appendix**486 **Tomographic method.**

The 3D anisotropic tomographic method is a classical two-step inversion procedure. From a practical point of view, the same algorithm (Tarantola & Valette 1982) is used for both steps. The forward problem can be generally written as:

$$\mathbf{d} = \mathbf{g}(\mathbf{m}) \quad (1)$$

where \mathbf{d} is the vector associated with the whole dataset, \mathbf{m} is the model parameter space, and \mathbf{g} is the functional relating \mathbf{d} and \mathbf{m} . The sensitivity kernels K are derived by first-order perturbation theory from equation (1). To generate an improved parameter estimate m_k starting from an initial parameter model m_0 , we use an iterative algorithm, as expressed in Tarantola & Valette (1982):

$$\mathbf{m}_k = \mathbf{m}_0 + C_m K_{k-1}^T M^{-1} [\mathbf{d} - \mathbf{g}(\mathbf{m}_{k-1}) + K_{k-1}(\mathbf{m}_{k-1} - \mathbf{m}_0)], \quad (2)$$

with

$$M = C_d + K_{k-1} C_m K_{k-1}^T$$

487 where K is the partial derivative matrix, $K_{k-1} = \frac{\partial \mathbf{g}(\mathbf{m}_{k-1})}{\partial \mathbf{m}_{k-1}}$. C_d is the covariance function of the data \mathbf{d} ,
 488 including the data uncertainties and possible correlations between the data, and C_m is the covariance
 489 function of the model parameters \mathbf{m} , limiting the accessible parameter space. The diagonal terms of
 490 the *a posteriori* covariance operator are also computed, in order to estimate the error bars on the final
 491 parameters.

In the first step, the higher mode phase velocities of the Rayleigh and Love waves are separately regionalized at different periods, to retrieve the different azimuthal terms of the local phase velocity. The azimuthal dependence due to the presence of a general slight anisotropic elastic medium is a homogeneous trigonometric polynomial up to degree 4ψ , where ψ denotes the azimuth of the horizontal wave vector, measured clockwise from the north. For a given point located at θ, ϕ and for a given angular frequency ω or period $T (= \frac{2\pi}{\omega})$, $V(\omega, \theta, \phi, \psi)$, the local phase velocity is defined as a perturbation of an isotropic or VTI (Vertical symmetry axis, Transversely Isotropic) reference velocity (denoted as $V_0(\omega, \theta, \phi)$), such that $V = V_0 + \delta V$. The velocity perturbation $\delta V(\omega, \theta, \phi, \psi)$ can be expanded as a Fourier series in ψ Smith & Dahlen (1973):

$$\delta V(\omega, \theta, \phi, \psi) = \frac{1}{2V_0} (A_0 + A_1 \cos 2\psi + A_2 \sin 2\psi + A_3 \cos 4\psi + A_4 \sin 4\psi). \quad (3)$$

492 The five local functions $A_n(\omega, \theta, \phi)$ depend on ω, θ, ϕ . Figure A1 (and similarly Figure A2) presents
 493 some global maps of A_0 (the local phase velocity) and A_1, A_2 (the $2-\psi$ azimuthal anisotropy), A_3, A_4
 494 (the $4-\psi$ azimuthal anisotropy) for the different Rayleigh wave (and Love wave) higher modes, at a

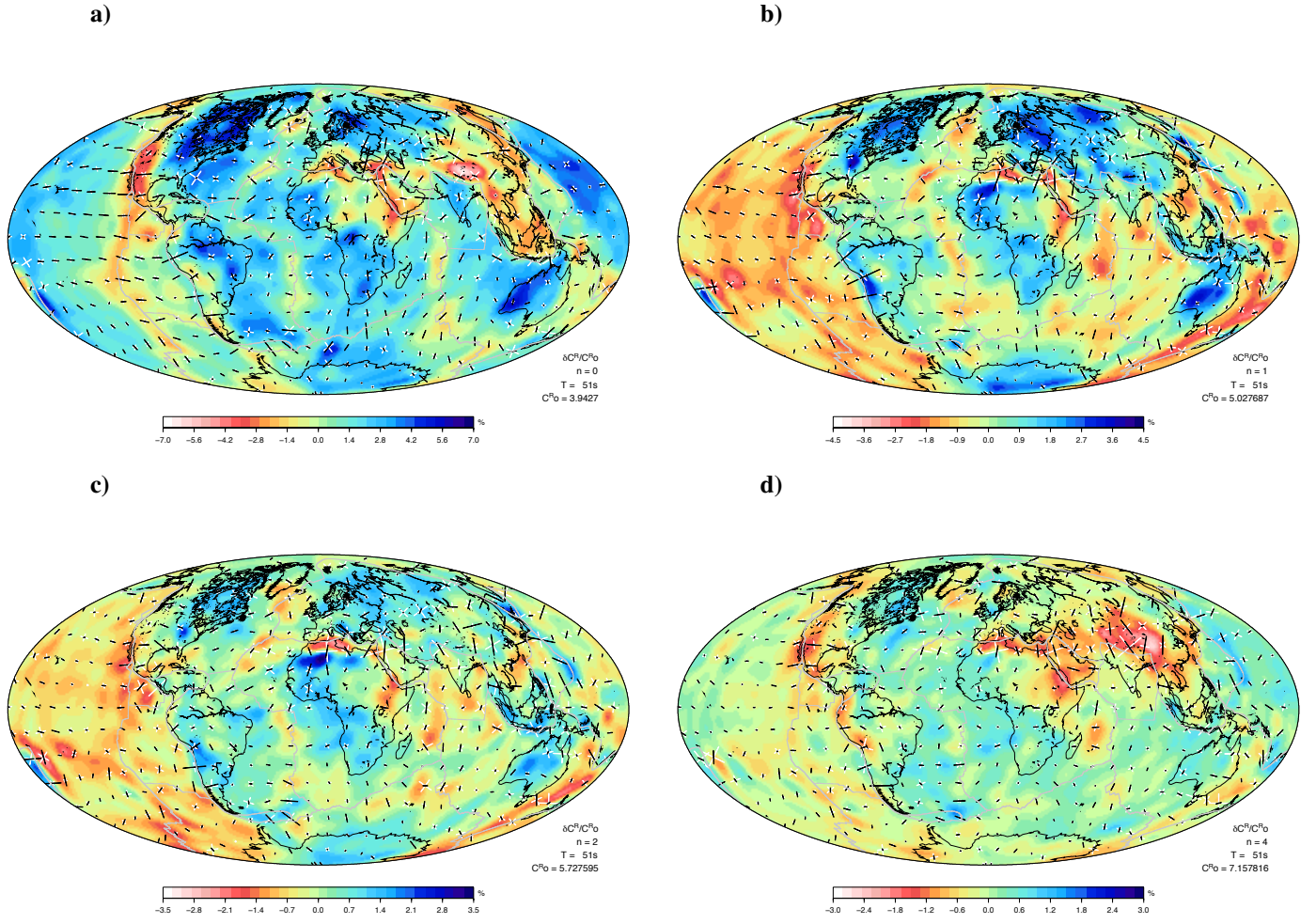


Figure A1. Phase velocity distributions of the different Rayleigh wave higher modes at the same period ($T = 51$ s). (a) $n=0$; (b) $n=1$; (c) $n=2$; (d) $n=4$. The black bars correspond to the maximum of $2 - \Psi$ azimuthal terms (azimuthal anisotropy), and their lengths are proportional to their amplitudes.

495 period of 51 s. All $A_n(\omega, \theta, \phi)$ are related to some of the 21 anisotropic parameters p_j that describe the
 496 medium (Montagner & Nataf 1986; Chen & Tromp 2007). These are inverted at every point (θ, ϕ) in
 497 the second step of the tomographic technique for retrieving an estimate of the whole parameter space
 498 at a given depth.

$$\delta V(\omega, \theta, \phi, \Psi) = \frac{1}{V_0} \sum_{j=1}^{n_p=21} \int_0^{r_a} K_j(\omega, r, \theta, \phi, \Psi) \delta m_j(r, \theta, \phi) dr$$

499 However, all of the parameters cannot be reliably retrieved. Figure A3 shows examples of the higher
 500 mode sensitivity kernels of the Rayleigh (Figure A3a) and Love waves (Figure A3b) with respect to
 501 the SV-wave and SH-wave velocities. The sensitivity of the Rayleigh waves to V_{SH} is small, whereas
 502 the Love wave higher modes can be sensitive to V_{SV} . The maximum sensitivity at a given period

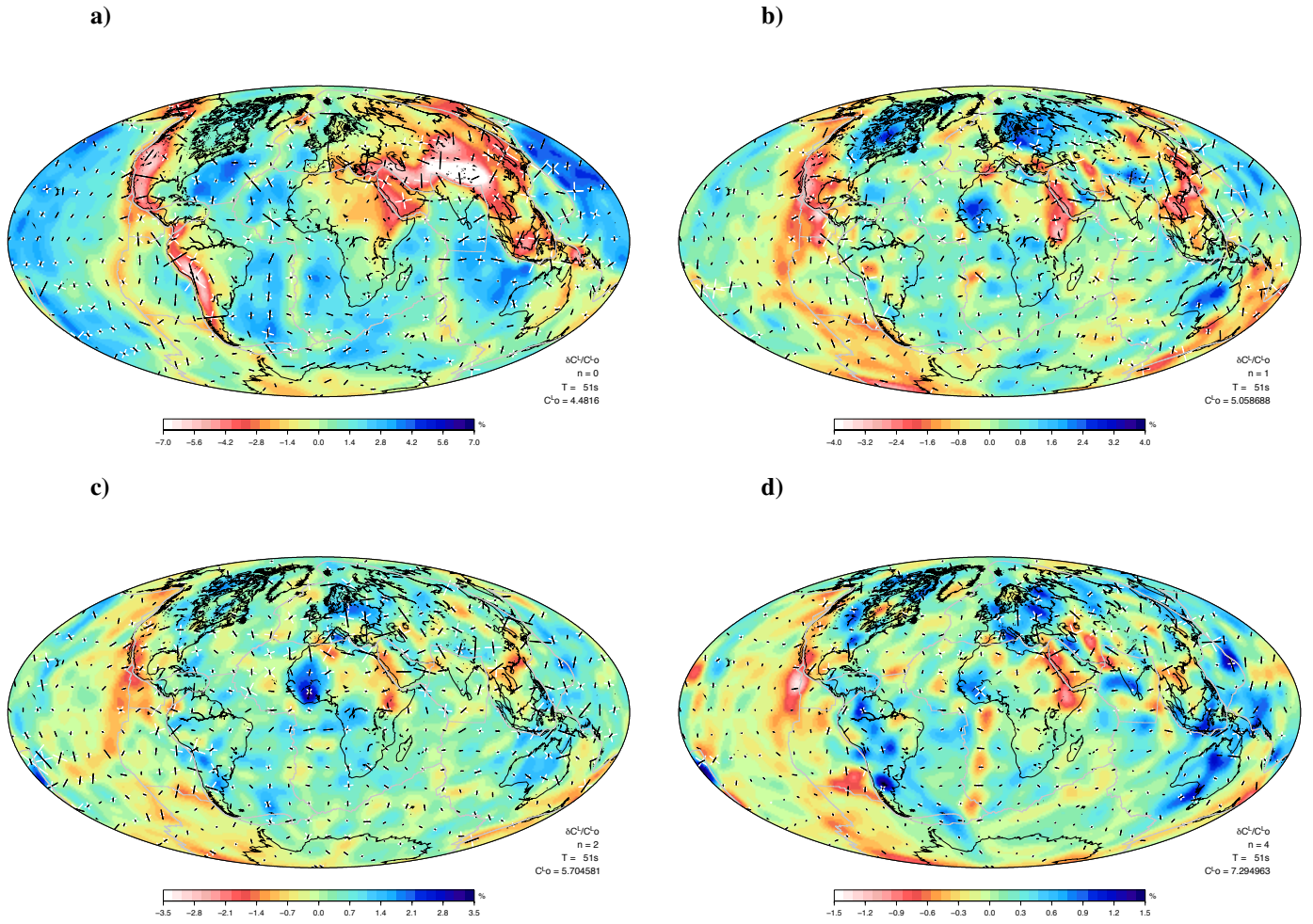


Figure A2. Phase velocity distributions of the different Love wave higher modes at the same period ($T = 51$ s). (a) $n=0$; (b) $n=1$; (c) $n=2$; (d) $n=4$.

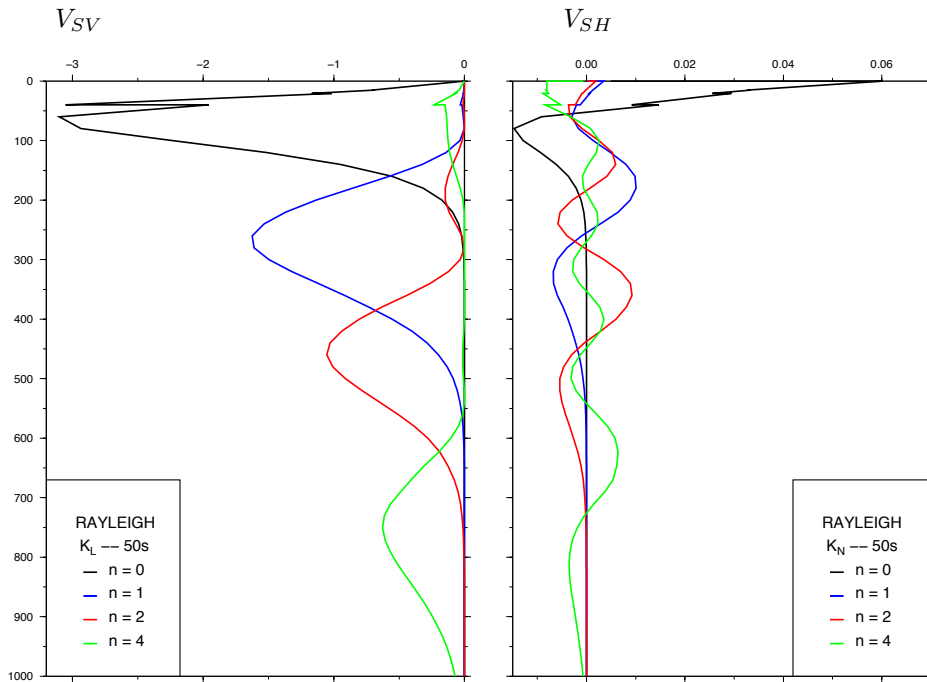
503 increases with the order n of the higher mode. For example, the 4th Rayleigh higher mode (Figure
 504 A3a) at 51 s has a maximum sensitivity in V_{SV} at 800 km in depth. These kernels show that higher
 505 modes make it possible to obtain good depth resolution down to 1200 km, well below the upper (410-
 506 660 km) and lower (660-900 km) transition zones, even though we only show the results down to 800
 507 km, which represents the main focus of the present study. Parameter G (and similarly parameter E)
 508 which expresses the $2 - \psi$ (and $4 - \psi$) azimuthal dependence of the SV-wave (and the SH-wave) has
 509 the same kernel as V_{SV} (and V_{SH}), and therefore it can also be recovered down to 1200 km in depth.
 510 V_{SV} , V_{SH} , G , Ψ_G , and E , Ψ_E are the most robust parameters that can be retrieved by the inversion
 511 at depth of the higher mode phase velocity distributions. Variations in ρ , V_{PV} and V_{PH} are scaled
 512 on these parameters, down to 410 km in depth Montagner & Tanimoto (1991), although they are free

513 below this depth. The local crustal structure has also been inverted due to the shortest periods of the
514 Rayleigh and Love wave dispersions in the datasets (Burgos 2013).

515 **Resolution tests**

516 We performed numerous synthetic tests to check the robustness of the inverted parameters ξ , G , Ψ_G , E , Ψ_E ,
517 by exploring different vertical correlation lengths, different weighing between the fundamental modes
518 and the overtones. Some of these examples are presented in Figure A4. We search for the depth res-
519 olution of the tomographic method by perturbing the depth profiles of V_{SV} and ξ at different depths
520 using delta functions convolved with a Gaussian filter (Figure A4, green lines). We show examples at
521 300 km and 700 km in depth. The important result is that in both cases, the perturbations are retrieved
522 at the correct depths. There is some spectral leakage towards other parameters V_{PH} , η , Φ , because we
523 invert for five parameters, whereas we only have two independent pieces of information in this case.
524 In addition, we can expect that in the real Earth, there should always be all kinds of anomalies for the
525 five parameters that will occur simultaneously.

a) Rayleigh



b) Love

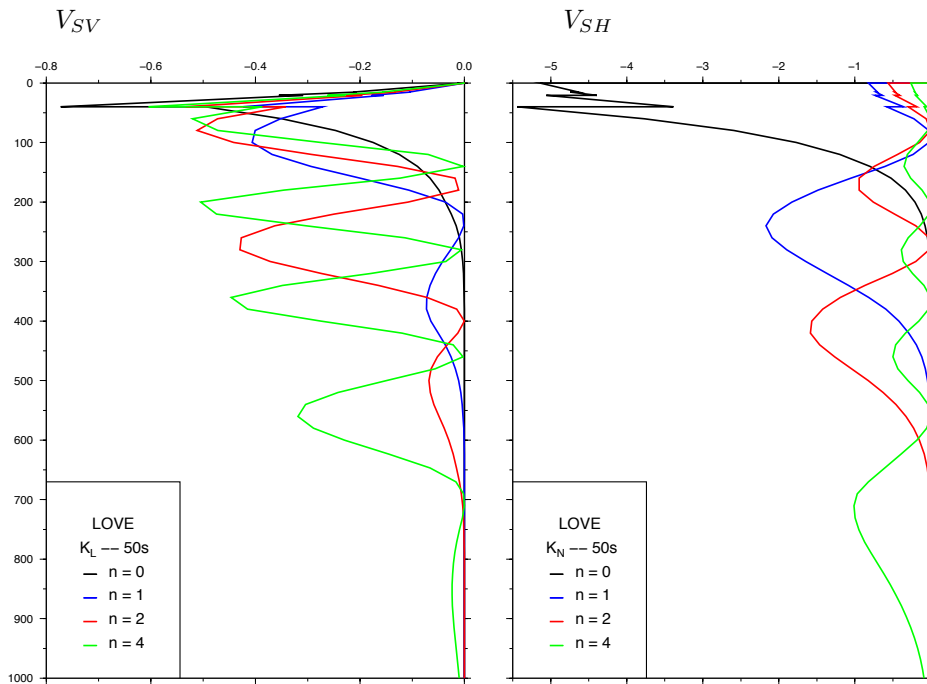
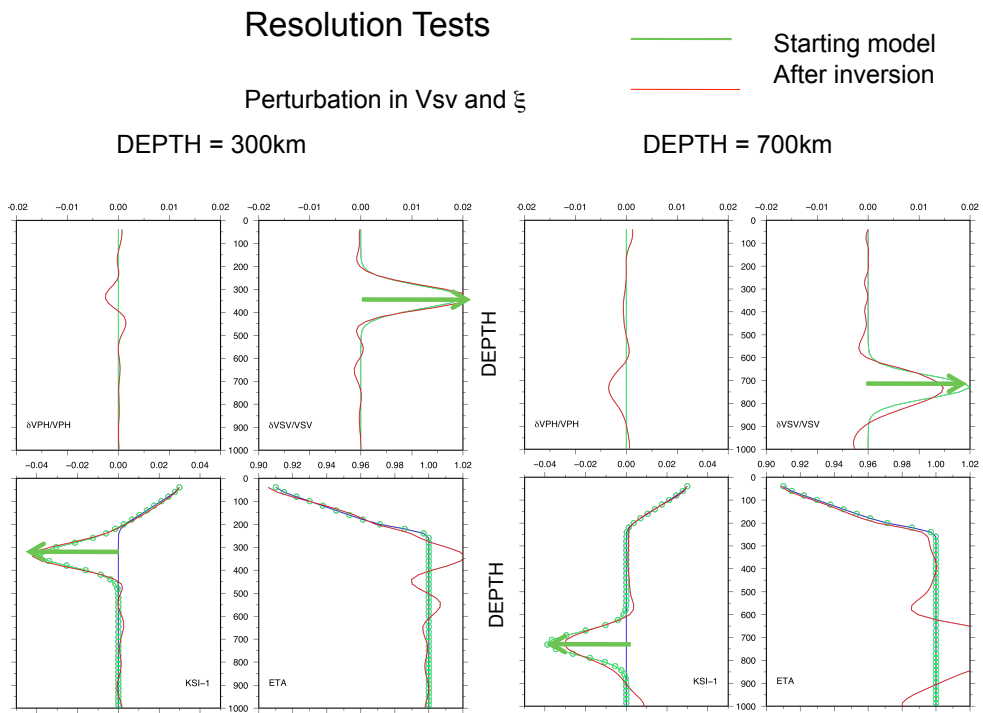


Figure A3. Phase velocity kernels of the different higher modes at the period $T = 51$ s. (a) Rayleigh waves. (b) Love waves. Left: L-kernel. Right: N-kernel. The elastic parameter L (and similarly N) is related to the V_{SV} velocity (and the V_{SH} velocity) by the relation $L = \rho V_{SV}^2$ (and $N = \rho V_{SH}^2$), where ρ is the density.



[h]

Figure A4. Resolution of the anisotropic parameters at the two different depths, of 300 km and 700 km.