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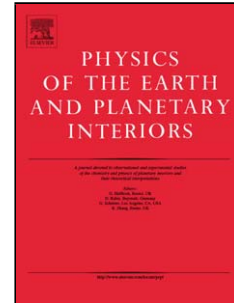
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# 1 Identifying global seismic anisotropy patterns by correlating 2 shear-wave splitting and surface wave data

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## 7 8 **Abstract**

9 We compare a global compilation of shear-wave splitting measurements with azimuthal seismic  
10 anisotropy parameters inferred from surface wave tomography. The currently available splitting  
11 dataset is taken from a novel comprehensive collection of available publications that is updated  
12 interactively online. The comparison between the two types of data is made by calculating  
13 predicted splitting parameters from the anisotropic tomography model. Comparing these  
14 predicted splitting parameters with the observed ones, we find a considerable correlation  
15 between the two datasets at global scale. This result is noteworthy, since such correlation did not  
16 seem to exist in previous studies. The spatial resolution associated with the two types of  
17 methods is rather different. While surface waves have good vertical resolution and poor lateral  
18 resolution of several hundreds of kilometers, SKS splitting measurements have good lateral, but  
19 poor vertical resolution. The correlation can be understood in light of recent propositions that  
20 anisotropy seen by SKS splitting constrains mostly the upper mantle, and therefore a similar  
21 depth region as surface waves. The correlation also confirms the generally good quality of the  
22 shear-wave measurements, as well as that of the anisotropic tomography model.

## 23 **1. Introduction**

24 Seismic waves are arguably the most powerful geophysical tools to investigate the Earth's deep  
25 interior. Anisotropy is one of the many material properties that affect seismic wave propagation. It  
26 manifests itself as a dependence of wave speed on direction. Such anisotropy can be caused by

27 structural variability, e.g. thin layers of alternating elastic properties (Backus, 1962) or fluid-filled  
28 cracks (Crampin, 1984; Kendall et al., 2006). A second origin which is widely accepted to  
29 dominate at mantle depths is the preferred orientation of anisotropic minerals by strain (e.g.,  
30 Nicolas and Christensen, 1987). Olivine plays a major role in the anisotropy of the Earth's mantle  
31 because it represents the primary constituent of the upper mantle and is more anisotropic than the  
32 other minerals. At upper mantle conditions olivine deforms by dislocation-creep (Nicolas and  
33 Christensen, 1987), and therefore aligns its main crystallographic axes with respect to structural  
34 directions (lineation, pole of the foliation), therefore producing a large-scale anisotropy detectable  
35 by seismic waves that is correlated with the geometry of strain in the upper mantle. Identifying the  
36 orientation of seismic anisotropy and its strength can thus help to constrain deformation at depth.  
37 Comparing seismic anisotropy with other observables such as gravitation and topography (Simons  
38 and van der Hilst, 2003) or the magnetic anomalies (Bokelmann and Wüstefeld, 2009; Wüstefeld  
39 et al., 2009) is helpful for understanding tectonic processes acting within the Earth, both, ancient  
40 ones recorded within the lithosphere, and more recent ones in the asthenosphere.

41 Hess (1964) was the first to measure azimuthal anisotropy from Pn travel-time variations beneath  
42 the northern Pacific and to interpret these variations in terms of mantle flow induced by sea-floor  
43 spreading. Since these pioneering works, a variety of techniques have been proposed for detecting  
44 seismic anisotropy. These include relative P-wave delay times (e.g., Babuska et al., 1984;  
45 Bokelmann, 2002), P-wave polarization (Bokelmann, 1995; Schulte-Pelkum et al., 2001; 2003;  
46 Fontaine et al., 2009), shear-wave splitting (for references see below) and radial and azimuthal  
47 anisotropic surface-wave tomography (e.g., Montagner, 1986, Nishimura and Forsyth, 1988;  
48 Montagner and Tanimoto, 1991).

49 The most direct evidence of seismic anisotropy is the splitting of shear-waves. Anisotropic media  
50 are indeed birefringent, splitting the incoming shear waves into two perpendicularly polarized  
51 waves that propagate at different velocities. At the Earth's surface, it is therefore possible to  
52 quantify anisotropy by measuring the two splitting parameters  $\delta t$  and  $\Phi$ . The difference in arrival  
53 times ( $\delta t$ ) between the two split shear waves (Figure 1) depends on both the anisotropy strength

54 and the length of the travel path through the anisotropic medium. The azimuth  $\Phi$  of the fast split  
55 shear wave polarization planes is related to the orientation of the anisotropic structure. Initially  
56 limited to direct S waves from local events (Ando and Ishikawa, 1980), the technique is now  
57 widely adopted for core-transiting phases such as SKS, SKKS, and PKS (e.g., Vinnik et al., 1984,  
58 1989; Silver and Chan, 1991; see Savage, 1999 and Fouch and Rondenay, 2006 for a review).  
59 Travelling on nearly vertical paths, teleseismic shear-waves provide measurements of anisotropy  
60 beneath a seismic station with a relatively good lateral resolution of few tens of kilometers  
61 (Fresnel-zone width) but with poor vertical resolution. The teleseismic shear-wave splitting  
62 technique has become popular in the last decades for performing anisotropy measurements that  
63 help revealing present or past mantle deformation processes. This technique has been widely  
64 applied in several geologic settings: Subduction zones (e.g., Levin et al., 2004; Margheriti et al.,  
65 2003; Nakajima and Hasegawa, 2004), rifts (e.g., Kendall, 1994; Gao et al., 1997; Walker et al.,  
66 2004; Kendall et al., 2005), hotspots (e.g., Barruol and Granet, 2002; Walker et al., 2001; 2005),  
67 oceanic islands (e.g., Behn et al., 2004; Fontaine et al., 2005, 2007), orogens (e.g., Barruol et al.,  
68 1998; Flesch et al., 2005) and stable continental environments (e.g., Fouch et al., 2000; Heintz and  
69 Kennett, 2005; Fouch and Rondenay, 2006; Assumpção et al. 2006).

70 Azimuthal anisotropy was initially deduced from surface waves by Forsyth (1975), then by  
71 Nishimura and Forsyth (1988) in Pacific and by Montagner and Tanimoto (1991) at global scale.  
72 The ever-growing dataset of seismic stations and recordings have allowed, over the years, to  
73 significantly improve spatial resolution of azimuthally anisotropic tomographic models (e.g.,  
74 Trampert and Woodhouse, 2003; Smith et al., 2004; Debayle et al., 2005; Maggi et al., 2006). By  
75 using the surface wave (Rayleigh and Love) frequency dependence on their penetration depth,  
76 surface wave tomography provides a relatively good vertical resolution of velocities and  
77 anisotropy down to the transition zone although the lateral resolution is relatively poor, generally  
78 not better than several hundreds of kilometres. In addition to the good vertical resolution, surface  
79 wave tomography has the advantage of providing a continuous and global-scale coverage of  
80 azimuthal anisotropy, except in poorly covered areas. This is in sharp contrast to the resolution of

81 teleseismic shear-wave splitting measurements which provide localized measures of anisotropy  
82 directly beneath a seismic station.

83 Given the constraints on deformation within the Earth that seismic anisotropy offers, there is much  
84 interest in understanding the origin of that anisotropy. Comparing observed fast directions with  
85 absolute plate motion directions is of interest (e.g., Vinnik et al., 1984; Silver and Chan, 1991;  
86 Montagner, 1994), since anisotropy in the asthenosphere, if caused by relative motion between  
87 plates and deeper interior of the Earth, should have fast directions parallel to plate motion. On the  
88 other hand, fossil anisotropy in the lithosphere does not need to show a correlation with current  
89 plate motion directions. It is also of interest to compare seismic anisotropy with predicted seismic  
90 anisotropy from flow calculations that have become available in recent years. That has been done  
91 at regional scale (Griot et al., 1998; Fouch et al., 2000; Becker et al., 2006), and also at global  
92 scale (Behn, et al., 2004; Becker et al., 2007; Conrad et al., 2007).

93 This study addresses the question whether shear-wave splitting observations and surface wave  
94 azimuthal anisotropy can be used interchangeably for the comparison with flow models. The  
95 different spatial resolution of the two types of data may result in significant differences: surface  
96 waves offer vertical resolution, while body waves offer lateral resolution. Furthermore, shear-wave  
97 splitting observations have mostly been made on continents (Figure 1), whereas surface wave  
98 models offer some resolution at all places of the Earth's surface, although poor ray coverage may  
99 result in weak constrains in some regions. A further complication for comparing the two datasets  
100 arises from the fact that the anisotropy is mainly characterised by two parameters, fast orientation  
101 and strength. A good coherence in one parameter does not necessarily means a good coherence in  
102 the other.

103 So far, there have been only few attempts to directly relate body-wave and surface wave  
104 anisotropy. Because the two techniques sample the anisotropy along vertical and horizontal  
105 propagation directions, respectively, the results cannot be compared directly. Montagner, et al.  
106 (2000) proposed to predict splitting parameters from a multi-layer anisotropic surface wave model.  
107 That study has suggested reasonable correlation at regional scale in some regions, but at global

108 scale there did not appear to be a clear correlation. The possibility was discussed that the spatial  
109 resolution of the two techniques is perhaps too different to allow realistic comparisons.

110 In this paper we follow the approach of Montagner et al. (2000), and we study the relation between  
111 anisotropy measures from shear-wave splitting and surface wave models at global scale, as well as  
112 for a few selected key regions. We will in fact show that there is indeed a correlation at global  
113 scale, and discuss its implications. In the first part of the paper, we present the new shear-wave  
114 splitting dataset. In the second part we present the prediction of splitting measurements obtained  
115 from recent surface wave tomography and in the third part, we compare the observed splitting data  
116 with the predicted parameters.

117

## 118 **2. World-wide shear-wave splitting results**

119 Over the past decades, a large number of shear-wave splitting studies have been published. Some  
120 of these studies have analyzed sets of globally distributed stations (e.g. Silver, 1996; Barruol and  
121 Hoffmann, 1999), while most focused on regional networks of stations, providing detailed  
122 knowledge on specific areas. Ever-larger temporary seismic arrays (e.g. USArray) and long-  
123 running permanent stations make the available shear-wave splitting dataset grow rapidly, and the  
124 need for a central depository for all available splitting measurements has been evoked (e.g., Fouch  
125 and Rondenay, 2006). Evans et al., (2006) present a shear wave splitting database, where they re-  
126 calculated available broadband stations. Our approach is different from that. In order to make all  
127 the published splitting data easily available in a homogeneous format, we developed an interactive  
128 database. The main features of this shear-wave splitting database are that the user may: i) Enter  
129 new splitting data via a web interface, ii) Search the database for a specific station or within a  
130 region, iii) Download the whole splitting parameters and corresponding bibliographic references  
131 and iv) Visualize the database, e.g. through GoogleEarth. Such an approach ensures that results of  
132 individual studies will be taken into account by the community, as users of splitting data will

133 always find the most “up-to-date” information. This database is available on the “Geosciences  
134 Montpellier” website at the address <http://www.gm.univ-montp2.fr/splitting/DB>.

135 As of June 2008, the database contains 2566 entries from 133 referenced studies (see Figure 2a).  
136 Note that the database is a collection of the work of various authors over many years, using a  
137 broad range of filters and possibly quality control factors. The global splitting delay times show an  
138 approximately log-normal distribution with most values between 0.4 and 1.4 sec and a maximum  
139 at 1 sec (Figure 3a, b). This distribution is in agreement with earlier studies, e.g. that of Silver  
140 (1996). Petrophysical analyses of upper mantle rocks sampled in orogenic or ophiolite peridotite  
141 massifs, or as kimberlite or basalt xenoliths suggest that such a mean delay time requires a 100 to  
142 150 km thick anisotropic layer (Mainprice and Silver 1993, Ben Ismail and Mainprice, 1998;  
143 Mainprice et al., 2000, Ben Ismail et al., 2001). We note that fast-axis orientations prefer E-W  
144 orientations, which are interesting in the light of speculations of uniform global plate motion  
145 relative to the underlying mantle (Doglioni, 1994; Steinberger and O’Connell, 1998; Doglioni et  
146 al., 2003). However, these statistics have to be considered with care, for two reasons: i) The ocean-  
147 continent discrepancy: Most splitting data originated from continents, where the azimuth may be  
148 dominated by frozen-in lithospheric anisotropy. Only a small amount of splitting data originated  
149 from oceanic environments where one could expect to observe present-day processes related to  
150 asthenospheric flow. ii) Seismic stations are not evenly distributed worldwide. Temporary  
151 experiments generally gather many instruments in a small area (e.g., Barruol et al., 1998; Walker  
152 et al., 2005; Flesch et al., 2005), which may cause a bias in statistical analyses. Furthermore, the  
153 different continents are not equally well sampled, and there is indeed an over-representation of  
154 North America and Western Europe in the dataset.

155

### 156 **3. Predicting shear-wave splitting parameters from surface wave data**

157 Montagner et al. (2000) developed a method to predict splitting parameters from surface-wave  
158 anisotropy results. Their approach is not unlike that of several earlier studies that discuss the effect



159 of multiple anisotropic layers on the effective splitting parameters (Silver and Savage, 1994;  
 160 Wolfe and Silver, 1998; Rumpker and Silver, 1998; Saltzer et al., 2000). The fast axis  $\Phi$  as well as  
 161 the delay time  $\delta t$  of a vertically propagating SKS phase that results from a (weak) anisotropic layer  
 162 is related to the elastic parameters  $G_s = C_{45}$ ,  $G_c = \frac{1}{2}(C_{55} - C_{44})$ , and  $L = \frac{1}{2}(C_{44} + C_{55})$  that  
 163 can be obtained from surface wave studies (Montagner, 1986). This can be generalized for  
 164 multiple layers. Montagner et al. (2000) have proposed a first-order expansion that is valid at  
 165 periods larger than 10 seconds, e.g. for SKS and SKKS waves. Under that assumption, splitting  
 166 delay and fast direction can be predicted by a simple integration along the (assumed vertical) ray  
 167 path, as

168

$$169 \quad \bar{d}t = \sqrt{\left(\int_0^a \sqrt{\frac{\rho}{L}} \frac{G_c}{L} dz\right)^2 + \left(\int_0^a \sqrt{\frac{\rho}{L}} \frac{G_s}{L} dz\right)^2} \quad (1)$$

$$\bar{\phi}_{fast} = \frac{1}{2} \arctan\left(\frac{\int_0^a \sqrt{\frac{\rho}{L}} \frac{G_s}{L} dz}{\int_0^a \sqrt{\frac{\rho}{L}} \frac{G_c}{L} dz}\right)$$

170 where  $a$  corresponds to the maximum depth extent of the tomographic model and  $\rho$  is the density  
 171 of each layer. In the following we will use the global tomographic model of Debayle et al. (2005).  
 172 Figure 2b shows predicted shear-wave splitting for that model, integrating from 500 km to 50 km  
 173 depth (thus excluding the negligible influence of the crust).

174

175 Debayle et al. (2005) have used observations of fundamental and higher-mode Rayleigh waves,  
 176 and among the 100779 ray paths, they include many short epicentre-station paths, which give  
 177 particularly strong constraints. The inversion has used a spatially variable cell size determined  
 178 using Voronoi tessellation, and the smallest cells have sizes of about 200 km. Since anisotropic  
 179 and isotropic variations often occur together, it is not likely that the very small variations are  
 180 resolved. Therefore, the tomographic inversion has also imposed a spatial smoothness constraint  
 181 via a Gaussian correlation function with standard deviation of 400 km. It is clear that resolution

182 must vary strongly with location along the Earth's surface, due to the variable ray coverage. In  
183 places where the coverage is particularly good, resolution will be better, but it is not quite clear  
184 what the resolution limit is. In principle, a Voronoi technique is well-adapted for optimally  
185 extracting information in this situation. In any case, a lower bound on resolution is imposed by the  
186 smoothness constraint corresponding to length scales of 400 km.

187 Figure 3c-f presents the statistical distribution of the predicted splitting parameters. The delay  
188 times range between 0 and 2 s, with a maximum of the distribution around 0.5 s, for both  
189 continental and oceanic environments. The predicted splitting delay times are thus systematically  
190 smaller delay than the observed ones (Figure 3a, b). Potential causes for this bias may be i) the  
191 damping that is necessarily applied during the surface-wave inversion, leading to an  
192 underestimation of parameters, ii) an underestimation of true lateral variations due to an  
193 insufficient lateral resolution of surface waves and/or iii) incompatibilities between the  
194 assumptions of anisotropy systems inherent to both methods. We choose to show the angular  
195 distribution separately for continental (Figure 3c, d) and oceanic regions (Figure 3e,f). For oceanic  
196 environments, the predicted splitting orientations display hardly any azimuthal preference on a  
197 global scale. However, for continental areas there are two maxima of predicted fast directions, a  
198 stronger one in N-S direction, and a weaker one in E-W direction. The N-S predicted fast  
199 directions appear primarily in Africa, north-east Asia, and Australia (Figure 2b), where station  
200 coverage for observed splitting measurements is poor. Such a discrepancy between predicted and  
201 observed fast split direction statistics could therefore be related to the extremely uneven station  
202 coverage present in the splitting database. However, at least for Africa, the N-S orientation is in  
203 accordance with recent surface wave tomographic models (Sebai et al.; 2006; Montagner et al.,  
204 2007). The E-W maximum is much less pronounced for surface waves than that for body waves,  
205 and it therefore does not serve to confirm this as a continental feature at the largest scale.

206

#### 207 4. Comparing observed and predicted SKS splitting

208 The two datasets we compare have rather different distribution on the Earth's surface. While the  
 209 splitting parameters are obtained at unevenly distributed seismic stations (Figure 2a), the predicted  
 210 splitting values from the surface wave model of Debayle et al. (2005) are evenly distributed on a  
 211  $1^\circ \times 1^\circ$  grid (Figure 2b). Comparing the two datasets therefore requires interpolation. Interpolating  
 212 sparse vectorial data as splitting measurements onto a regular grid is possible (Bird and Li, 1996;  
 213 Lucente et al., 2006), but to avoid stability problems due to the different spatial support, we prefer  
 214 to follow the approach proposed by Montagner et al. (2000), i.e., we linearly interpolate the  
 215 regular grid of the predicted anisotropy parameters to the location of each splitting measurement.

216 The correlation between the observed and predicted datasets is then obtained by using an adapted  
 217 form of the coherence function initially defined by Griot et al. (1998). This mathematical function  
 218 quantifies the correlation between two vectorial data sets on a sphere (co-latitude  $\Theta$ , longitude  $\Psi$ )  
 219 by taking into account the two anisotropy parameter pairs  $(\Phi_1, \delta t_1; \Phi_2, \delta t_2)$ . Since each parameter  
 220 pair corresponds to a phase and amplitude, a natural form of the coherence between two  
 221 continuous fields on the sphere is

$$222 \quad C(\alpha) = \frac{\iint_{\Theta \Psi} \left\{ dt_1(\Theta, \Psi) * \sin \Theta * dt_2(\Theta, \Psi) * \sin \Theta * \exp \left( - \frac{(\phi_1(\Theta, \Psi) - \phi_2(\Theta, \Psi) + \alpha)^2}{2D_{corr}^2} \right) \right\} d\Theta d\Psi}{\sqrt{\iint_{\Theta \Psi} dt_1(\Theta, \Psi)^2 * \sin^2 \Theta d\Theta d\Psi} \sqrt{\iint_{\Theta \Psi} dt_2(\Theta, \Psi)^2 * \sin^2 \Theta d\Theta d\Psi}}$$

223 (2)

224 The phases (fast directions) enter through an exponential form that is normalized by a “correlation  
 225 factor”  $D_{corr}$ . This formula does not require an alignment of the two sets of fast axes. The test  
 226 orientation  $\alpha$  can vary between  $-90^\circ$  and  $+90^\circ$  and allows testing for an angular bias between the  
 227 two datasets. The normalization in the conventional fashion ensures that the coherence varies  
 228 between 0 (uncorrelated) to 1 (perfect correlation). An anti-correlation between the fast axes  
 229 would show up as a peak at  $\pm 90^\circ$ .

230 If  $C(\alpha)$  is represented by a horizontal line, the two models are uncorrelated. A Gaussian shape of  
231  $C(\alpha)$  indicates a certain coherence, with the location of the peak corresponding to the mean  
232 rotation between the models. Figure 4a illustrates this by comparing the  $1^\circ \times 1^\circ$  predicted splitting  
233 dataset with itself. As expected, this results in a Gaussian distribution with a maximum value of 1  
234 oriented at  $0^\circ$ . Figure 4b shows the coherence of the predicted splitting data on a  $1^\circ$  grid with that  
235 of a  $10^\circ$  grid. Maximum values are considerably smaller than 1 now, and the distributions are  
236 much wider, reflecting the added variance from the interpolation using the  $10^\circ$  grid. In fact, there is  
237 much variation of predicted fast directions in Figure 2b, at scales below  $10^\circ$ .

238

#### 239 **4.1. Global correlation of splitting parameters**

240 Montagner et al. (2000) compared splitting parameters as predicted by a global anisotropic  
241 tomography model (Montagner and Tanimoto, 1991) with the shear-wave splitting results  
242 compiled by Silver (1996). The results suggested a fairly good coherence at regional scale (for  
243 instance in the Western United States and in Central Asia), but no clear coherence at global scale.  
244 The authors have attributed the latter to the different depth and lateral sensitivity in both datasets.  
245 Surface waves are best resolved in oceanic environment, where large-scale structures are expected  
246 but where station coverage and therefore shear-wave splitting measurements are sparsely  
247 distributed in that environment. In contrast, shear-wave splitting has a relatively good lateral  
248 resolution and may evidence short scale variations in upper mantle structures beneath the  
249 continents that are hardly visible from surface wave analyses. The authors also suggested that the  
250 usually-performed single-layer inversion for splitting parameters may represent an  
251 oversimplification that contributes to the missing global correlation.

252 Figure 4c and d show the coherence of predicted splitting from the model of Debayle et al. (2005),  
253 integrated between 50-500km depth, with the splitting database, taking into account the entire  
254 database using gridding sizes of  $1^\circ$  and  $10^\circ$ , for the correlation lengths used previously ( $D_{\text{corr}} = 5,$   
255  $10$  or  $20^\circ$ ). The coherence is slightly broader for the  $10^\circ$  grid (Figure 4d) than for the  $1^\circ$  grid

256 (Figure 4c). However, both comparisons give a rather good correlation of available splitting data  
257 and anisotropic surface wave model, with a maximum centred near  $0^\circ$  misfit. This indicates that  
258 the available splitting database sufficiently covers the globe, as it is insensitive to the resolution of  
259 the surface wave model.

260 In the former study, Montagner et al. (2000) used the anisotropic surface wave model of  
261 Montagner and Tanimoto (1991) and the splitting data collection of Silver (1996). The global  
262 coherence was found to show a local minimum at  $0^\circ$  misfit, with local maxima on both sides. We  
263 were able to roughly reproduce this behaviour by incorporating only the splitting measurements  
264 available before 1997 (Figure 4e). Montagner et al. (2000) concluded that there is no global  
265 correlation. In fact, Figures 4d-f suggest that the absence of correlation in that study was  
266 apparently an effect of insufficient splitting data coverage and limited spatial resolution in  
267 available surface wave models at the time. Since Fresnel zones of the commonly used SKS waves  
268 are of the order of 100 km at upper mantle depths, and since splitting parameters vary at these  
269 scale length (Figure 2a), averaging azimuthal anisotropy at scale larger than  $10^\circ$  introduce a too  
270 large smoothing to preserve a good coherence between the body wave pattern and the large-scale  
271 surface wave pattern.

272 The higher correlation of the new global set (Figure 4c) in comparison to sub-dataset of Silver  
273 (1996; Figure 4e) may on the other hand reflect the numerous regional studies performed after  
274 1996, that concentrate a large number of stations in relatively small areas, e.g. in the Eifel region  
275 (Walker et al., 2005), and the Pyrenees (Barruol et al., 1998). This leads to regional biases. The  
276 difference suggests that we should try to eliminate the regional biases if possible, or at least  
277 eliminate the effect of the spatially varying data density where we have data. So far, all data points  
278 enter equally into the coherence, and local studies with dense station coverage will thus have a  
279 stronger effect on the coherence than sparsely instrumented areas. We address this problem by  
280 gridding the splitting data. However, standard interpolation procedures are inappropriate when  
281 interpolating sparse directional data on a sphere (Bird and Li., 1996). One approach to solve this  
282 problem may be by using Voronoi cells, similar to the method of Debayle et al. (2005) or Larmat

283 et al. (2006). In this study, however, we follow a different approach: To account for both the  
 284 amplitude and orientation of anisotropy we separate amplitude (delay time) and orientation (fast  
 285 direction). The delay time is gridded using a standard cubic spline interpolation to give a map of  
 286 gridded delay times  $T$ . The directional component of anisotropy is computed using an algorithm  
 287 that has already been successfully applied to global stress data (Müller et al., 2003). The gridding  
 288 algorithm is based on Watson (1985) and extended by a distance-weighting (Wehrle, 1998). For a  
 289 series of unit vectors  $\vec{y}$ , given at positions  $\vec{x}$ , Watson (1985) proposes a technique that searches  
 290 for a ‘predictor’ function  $\vec{f}(\vec{x})$  that describes the vector field parallel (or anti-parallel) to  $\vec{y}$ . This  
 291 is achieved by maximising the (quadratic) scalar product  $P = \langle \vec{f}(\vec{x}_i)^T \vec{y}_i \rangle^2$ . Furthermore, to be  
 292 smooth, the (quadratic) scalar product of two neighbouring functions  $N = \langle \vec{f}(\vec{x}_i)^T \vec{y}_j \rangle^2$  must be  
 293 maximum as well. The resulting gridded vector field is thus the maximum of the following  
 294 function  $K(f)$ :

$$295 \quad K(f) = \sum_{i=1}^n \left( Pd(\vec{x}_i, \vec{x}_i) + \lambda \sum_{j=1, j \neq i}^n Nd(\vec{x}_j, \vec{x}_i) \right) \quad (3)$$

296 As mentioned, the P-term achieves optimum fitting of the data, the N-term controls the  
 297 smoothness of the resulting field. The weighting factor  $\lambda$  lambda thus controls the relative  
 298 influence of the data fidelity and smoothness, respectively. The function  $d(\vec{x}_i, \vec{x}_j)$  expresses the  
 299 distance weighting of point  $\vec{x}_j$  at  $\vec{x}_i$  within a search radius  $R$  (Wehrle, 1998). The interpolated  
 300 splitting “vector field” is then the linear combination of the gridded splitting amplitude and  $K(f)$ ,  
 301 calculated at the locations  $\vec{s}_i$ . These coincide with the grid points of the tomographic model and  
 302 thus the location of the predicted splitting parameters.

$$303 \quad V = T|_{\vec{s}} * K(f)|_{\vec{s}} \quad (4)$$

304 Figure 5 shows an example result of that gridding for  $R = 600$  km and  $\lambda = 5$  on a  $3^\circ \times 3^\circ$  grid. Null  
 305 measurements, though important to constrain fast directions at individual stations (Wüstefeld and  
 306 Bokelmann, 2007), are not taken into account.

307 Similar to Figure 4, we calculate the coherence functions of the predicted splitting with these  
308 smoothed global datasets, to ascertain that the correlation between observed and predicted splitting  
309 as described in the previous section, is not an artefact of the rather heterogeneous data density. If  
310 the correlation is real, this helps to define the inherent lateral resolvability of anisotropy by the  
311 used surface-wave model.

312 The results for various parameter combinations of  $R$  and  $\lambda$  are presented in Figure 6a ( $D_{\text{corr}}$  is fixed  
313 to  $10^\circ$ ). For all tests, there is a maximum near  $0^\circ$ , suggesting that there is indeed a correlation also  
314 for the gridded data, similar to the global correlation found previously (Figure 4). The correlation  
315 between the two datasets is therefore not an artefact of the uneven sampling.

316 Figure 6b shows the dependence of the coherence value at  $0^\circ$  on search radius for various  $\lambda$ .  
317 Interestingly, the coherence decreases clearly below  $R = 400\text{km}$  while it stays more or less  
318 constant between 400 and 1000 km. Below  $R = 400\text{km}$ , the coherence shows a weak dependence  
319 on smoothing parameter  $\lambda$ . This is as expected, since the effect of  $\lambda$  must be somewhat similar to  
320 that of the search radius. Thus, the decrease of coherence below  $R = 400\text{km}$  also manifests itself in  
321 the  $\lambda$  values. A peak in coherence is obtained for strong smoothing with  $\lambda = 10$  and 50 at search  
322 radii of  $R = 400$  km. Moreover, a plateau of relatively good coherence is found for  $\lambda = 5$  and  
323  $R$  between 600 and 700 km.

324 Figures 6c and 6d are similar to 6a and 6b, but the order of each parameter ( $\Phi_1, \delta t_1; \Phi_2, \delta t_2$ ) has  
325 been randomised. In such random test any spatial coherence should be eliminated, unless there is  
326 an inherent correlation, due to a non-random data distribution. Comparing Figure 6c with Figure  
327 6a we note that coherence indeed drops sharply, indicating that the correlation observed in Figure  
328 6a is indeed real. The weak residual coherence in Figure 3b thus shows the artefact of the non-  
329 uniform distribution of splitting fast directions.

#### 330 **4.2. Regional correlation of splitting parameters**

331 The previous section showed that anisotropy as determined from surface waves and those from  
332 body waves show similar results on a global scale, though both methods sample the Earth at

333 different scales. In this section we discuss several regional subsets of different tectonic  
334 background. Montagner et al. (2000) point out that several regional subsets have a rather good  
335 coherence despite the poor global coherence obtained in their earlier study. They relate that  
336 partially to an increased resolution of regional tomographic models used. We limit our comparison  
337 here to regional subsets of the global model. Regional tomographic models might provide higher  
338 resolution, yet a detailed comparison of all available models would be beyond the scope of this  
339 study. Figure 7 illustrates the coherence function for several regional subsets in different tectonic  
340 settings (subduction, rifts, volcanic islands, cratons). See Figures 2 and 5 for locations. Generally,  
341 the coherence is very good, with a clear maximum centred close to  $0^\circ$  misfit. Only the absolute  
342 value of that maximum varies between the regions. We attribute this to differences in the observed  
343 and predicted delay times, as we will discuss below.

344 Montagner et al. (2002) selected Western United States and Central Asia for closer discussion. For  
345 consistency and comparison with that study we will briefly discuss the results of our new  
346 coherence study, and also present results for other selected regions and tectonic settings.  
347 Additional maps of test regions can be found in the Appendix. This section shall provide an  
348 overview on coherences between the two datasets. For detailed interpretation of the anisotropy in  
349 the selected regions we kindly refer to the according publications.

#### 350 4.2.1. The Andes

351 Several studies analysed shear-wave splitting in the Andes and adjacent Nazca plate. Polet et al.  
352 (2000) argue that most significant contribution to splitting occurs below the slab, with a complex,  
353 spatially varying anisotropy pattern. Russo and Silver (1994) and Helffrich et al. (2002) propose  
354 that the large-scale Pacific mantle flow pattern is deviated around the South American continent by  
355 roll-back of the Nazca slab.

356 The coherence between predicted and interpolated observed splitting is variable in this area, with  
357 poles at  $-70^\circ$ ,  $+20^\circ$  and  $+60^\circ$  (Figure 7a, Figure A1). This behaviour renders the complex  
358 interaction, sometimes at small scale, between subduction of oceanic plate and mountain range



359 formation, both of which may cause different anisotropy orientations (Vauchez and Nicolas, 1991;  
360 Barruol et al., 1998; Kneller et al., 2007).

#### 361 4.2.2. The East Pacific Rise

362 The coherence between predicted and interpolated observation of splitting is striking for stations in  
363 the vicinity of the East Pacific Rise (Figure 7b; Figure A2). These measurements come from the  
364 MELT and GLIMPSE Ocean Bottom Seismometers experiments (Wolfe and Solomon, 1999;  
365 Harmon et al., 2004). They proposed Lattice Preferred Orientation (LPO) of anisotropic minerals  
366 parallel to mantle flow trending normal to the spreading axis as main cause for splitting, as well as  
367 a deeper contribution of an asthenospheric return flow from the South Pacific superswell. This  
368 agreement between the fast split directions and the surface wave anisotropy is not restricted to the  
369 ridge itself but is extending further West in the French Polynesia domain where Fontaine et al.  
370 (2007) evidenced a clear signature of the asthenospheric flow in the SKS splitting observations  
371 that also dominates within the surface wave tomographies, both exhibiting dominant fast directions  
372 parallel to the plate motion direction. Further south, the EPR region also contains station RPN,  
373 located on the Easter Island Hot Spot, for which a multilayer case has been proposed (Fontaine et  
374 al., 2007). The SKS anisotropy pattern observed at this station is clearly not the result of the ridge  
375 spreading alone and may likely result from the combination of spreading-related mantle flow and  
376 hotspot-related mantle upwelling. South and east of station RPN, although both the station  
377 coverage and resolution of the surface wave model is too small to confidently relate this behaviour  
378 to mantle upwelling, the predicted and interpolated splitting coincide rather well, whereas towards  
379 the north, the two models are almost perpendicular. This can be seen in Figure 7b as a smaller  
380 peak at  $-65^\circ$ .

#### 381 4.2.3. North America

382 The understanding of regional tectonics in North America is continuously improved (e.g.,  
383 Humphreys and Coblenz, 2007). Seismic anisotropy beneath North America has been addressed  
384 by numerous authors (see references in Walker et al., 2004; Fouch and Rondenay, 2006,  
385 Bokelmann and Wüstefeld, 2009). The pertaining mantle flow is mainly controlled by the various

386 tectonic settings ranging from cratons in the north (e.g., Bokelmann, 2002a, b), mountain ranges in  
387 the east (e.g., Barruol et al., 1997; Fouch et al., 2000), and continental hotspot (e.g., Waite et al.,  
388 2005) and subduction in the west that could interfere with strike-slip tectonics to produce a  
389 complex toroidal mantle flow pattern (e.g., Xue and Allen, 2007; Zandt and Humphreys, 2008).  
390 The splitting dataset presented in this study indicates good correlation with the splitting predicted  
391 from surface wave azimuthal anisotropy in North America (Figure 7c, Figure A3). Montagner et  
392 al. (2000) observed however a rotation of  $+30^\circ$  between observed and predicted splitting in the  
393 Western US (Figure 8a). In this region, the coherence of the new dataset has improved (Figure 7d),  
394 likely due to the larger amount of splitting data (109 measurements today versus 47 available in  
395 1996), but perhaps also due to the more recent tomographic model used in this study that may  
396 provide an improved lateral resolution. The resulting misfit is reduced to a slight rotation of  $-10^\circ$   
397 for all of North America. In Western North America the angular misfit is precisely  $0^\circ$ . This  
398 suggests that surface waves and shear wave splitting experience the same anisotropic structures  
399 (Figure 8a) and therefore, that the apparent complexity of the crustal tectonics in this region may  
400 be overlying a mantle structure characterized by a simpler pattern at longer wavelength. It thus  
401 appears to be consistent to invert the two datasets together as has been done recently by Marone  
402 and Romanowicz (2007).

#### 403 4.2.4. Kaapvaal Craton

404 Several studies analysed the Kaapvaal upper mantle anisotropy by using either SKS wave splitting  
405 (e.g., Vinnik et al., 1996; Barruol and Ben Ismail, 2001; Silver et al., 2001; Fouch et al., 2004) or  
406 the information brought to the surface by kimberlite nodules (e.g., Ben Ismail et al., 2001). Silver  
407 et al. (2001) and Gao et al. (2002) point out that the fast orientations do not correlate with plate  
408 motion directions of Gripp and Gordon (2002). The fast orientation correlates however very well  
409 with regional geologic trends, suggesting lithospheric origins of anisotropy. Furthermore, a high-  
410 density array in the Kimberley region found small-scale variations in shear wave splitting  
411 parameters, which can only be explained by lithospheric origins of anisotropy (Fouch et al., 2004).

412 In global tomographic models, one of the most intriguing features is the African Superswell (e.g.,  
413 Nyblade and Robinson, 1994; Ritsema and van Heijst, 2000; Conrad and Gurnis, 2003), which is  
414 an ongoing, large-scale, northward oriented mantle upwelling. This upwelling can provide an  
415 explanation for the predicted splitting (Figure 2b, 8b) particularly because of the absence of  
416 predicted splitting beneath the Namibia-Botswana border region (Figure 8b), since vertically  
417 oriented anisotropy is not included in the methods used here. The presence of two peaks in the  
418 regional coherence function (Figure 7e) may indicate that the anisotropy observed by tomographic  
419 studies is not only related to the ongoing mantle upwelling, but also related to ancient lithospheric  
420 anisotropy, at least in the upper mantle regions.

#### 421 4.2.5. The East African Rift Zone

422 Rifting along the East African Rift Valley (EAR) has been analysed at mantle to crustal scales by  
423 various studies (see Kendall et al., 2006 for a review). Collectively, the results support a model for  
424 magma assisted rifting in Ethiopia with an increase in melt production towards the Red Sea, which  
425 is strongly supported by the prediction of partial melt on the basis of P- and S-wave tomography  
426 (Bastow et al., 2005, 2008). In Tanzania and Kenya, fast split shear waves parallel the E and W  
427 arms of the EAR that encircle the Tanzanian craton (Gao et al., 1997; Walker et al., 2004).

428 The coherence between observed and predicted splitting is rather strong with a maximum at  $0^\circ$   
429 deviation (Figure 7f, Figure A4). The strong coherence is mainly due to high correlation of both  
430 direction and amplitude of splitting between the predicted and interpolated datasets beneath Afar  
431 (Figure A4). Kendall et al. (2006) proposed a suite of different causes of anisotropy beneath the  
432 East African rift system. The main cause of anisotropy is attributed to thin melt pockets directly  
433 beneath the rift zone, oriented parallel to the rift axis. Away from the rift axis, the lithospheric  
434 LPO is likely related to pre-existing fabric, while beneath the lithosphere the LPO could be due to  
435 viscous coupling between the base of the lithosphere and large-scale mantle upwelling. This model  
436 explains the rift-parallel fast splitting orientations in the rift valley. Such explanation leaves the  
437 good correlation somewhat surprising, since global surface waves are not expected to resolve that  
438 feature. The splitting delay times predicted in this area are relatively low, which may be due to two

439 perpendicular anisotropic orientations (see discussion for Australia). An alternative explanation  
440 might be that the mantle upwelling beneath the rift causes vertically oriented fast axis in the  
441 mantle. These do not contribute to the observed and the predicted splitting.

#### 442 4.2.6. *Central Asia*

443 In Central Asia (Figure 7g, Figure A5), more than 250 measurements are now available, in  
444 comparison with the 33 available before 1997. The new model seems to be less coherent (Figure  
445 7g) than the previous one, as discussed in Montagner (2000), and it is characterized by the  
446 presence of two peaks in the coherence function at  $+15^\circ$  and  $-70^\circ$ . Montagner et al. (2000) used a  
447 local high-resolution tomographic model (Griot et al., 1998) for calculating the predicted splitting.  
448 The two compared datasets seem to be rotated by  $20^\circ$ . The main region of discrepancy is the  
449 Tibetan Plateau, while coherence is good for the Tien Shan basin in the North-West and the North  
450 China craton, in the North-East of Figure A5). In general, discrepancies between the global  
451 (Montagner and Tanimoto, 1991) and local (Griot et al, 1998) tomographic datasets highlight the  
452 importance of local high-resolution tomography that may be sensitive to shorter wavelength  
453 structures. This regional analysis also suggests that more data do not necessarily imply a better  
454 coherence since local studies involving a large number of stations over a limited space may  
455 strongly bias the general coherence function at the regional and global scale. They conclude that  
456 future approaches should consider the effect of different data density in greater detail. New  
457 azimuthal anisotropy tomographic models of Central Asia with an enhanced lateral resolution are  
458 presently issuing. Consequently time is ripe for a new quantitative comparison between SKS  
459 splitting data and tomographic models, but it is beyond the scope of this paper.

460

#### 461 4.2.7 *Australia*

462 The coherence in Australia is rather good (Figure 7h, A6). Given the complex anisotropy  
463 described by other studies of this area, this is somewhat surprising. Debayle et al. (2005) found  
464 that Australia is the only continent where azimuthal anisotropy (inferred from surface waves)

465 correlates well with present day absolute plate motion. This correlation is most significant at  
466 depths between 150 and 300km. In Australia many stations are observed as isotropic, i.e. Nulls  
467 from a broad range of backazimuths. Heintz and Kennett (2005, 2006) point out that this apparent  
468 isotropy at these stations is also consistent with two layers of mutually perpendicular anisotropy  
469 orientations (Silver and Savage, 1994; Barruol and Hoffmann, 1999).

470 The predicted shear wave splitting contains a few regions of negligible splitting amplitude (central  
471 Northern Territory, just south of Western Australia, and in Central New South Wales; see Figure  
472 A6). In our approach, the predicted splitting is calculated from 19 layers, each 25 km thick. The  
473 correlation between non-Null observations and predicted splitting is very good, indicating that  
474 such gradual change in anisotropy orientation with depth mimics the non-Null splitting  
475 observations well. This in turn indicates that assuming a two layer case may be an  
476 oversimplification.

## 477 5. Discussion

478 Generally, the resolution of global anisotropic surface wave models is thought to not be better than  
479 a thousand kilometres. For an increasing number of regions, recent high-resolution regional  
480 anisotropic tomographic models are available, and future studies should address their correlation  
481 with regional splitting data in more detail. For the global model, the dependence of correlation on  
482 search radius in Figure 6b may suggest that resolution capability is somewhat better than the  
483 initially thought thousand kilometres. This is true at least for the parts of the Earth that contain the  
484 most seismological stations. These areas correspond to the seismically best-resolved portions on  
485 Earth, obviously for shear wave splitting but also in surface wave data due to the higher density of  
486 crossing ray paths in those areas. The use of Voronoi cells allows the tomographic model of  
487 Debayle et al. (2005) to reflect spatially varying resolutions. The drop-off in correlation below 400  
488 km is interesting, since we would be able to see that effect only if there is resolution of surface  
489 wave data for those best-resolved parts of the Earth down to 400 km length scales, which is  
490 considerably better than we expected. Whether or not this is the correct explanation for the decay  
491 below 400 km, this resolution study is very encouraging, since it indicates lateral surface wave

492 resolution somewhat below a thousand kilometres for the well-resolved portion of the Earth. The  
493 resolution will be considerably weaker in less well-covered regions. A recent study of the  
494 Canadian shield (Bokermann and Wüstefeld, 2009) has suggested that surface-wave anisotropy  
495 models do not yet have a similar resolution there, which would be required to distinguish inner-  
496 cratonic variations of anisotropy.

497 This will probably change somewhat in the next years, with the availability of massive regional  
498 datasets. Also methodological developments such as simultaneous inversions of splitting and  
499 surface wave data are quite useful (e.g., Marone and Romanowicz, 2007).

500 A limitation of our method is that no Null stations, i.e. apparently isotropic stations, are included.  
501 The explanation of such Null stations is not yet completely understood as ever more evidence  
502 suggests that at least the upper part of the mantle is anisotropic. This evidence is based on  
503 seismological observations as presented in this paper and also on mineralogical studies (e.g., Ben  
504 Ismail and Mainprice, 1998; Mainprice et al., 2000; Holtzman et al., 2003). Nulls stations do not  
505 necessarily imply pervasive upper mantle isotropy, but may for example be caused by two  
506 anisotropic layers with mutually perpendicular fast orientations (e.g., Silver and Savage, 1994;  
507 Heintz and Kennett, 2006). Null stations may also not imply large scale upper mantle isotropy. For  
508 instance, P-polarization analysis at station PPT in Tahiti Island (Fontaine et al., 2009) clearly  
509 demonstrates that the isotropy deduced from SKS splitting, located vertically beneath the station,  
510 has a restricted lateral extent, implying that such a short-scale upper mantle structure can hardly be  
511 visible in surface wave tomographies.

512 The database contains only a limited amount of multi-layer cases. For the bulk of data, the  
513 observed (apparent) splitting parameters reflect the vertical integration of the different layers of the  
514 surface wave anisotropic model. Although the model is divided in 19 layers between 40 and 500  
515 km depth, the radial correlation length limits the actual number of independent layers to 4 or 5.  
516 These layers can in turn display effectively the same azimuth and so the real number of layers with  
517 different orientations (in complex areas) may be at the most two or three. However, it is important  
518 to note that in many areas these azimuthal variations and this stratification of anisotropy are real

519 and significant, although they cannot directly be inferred from shear wave splitting (Silver and  
520 Savage, 1994; Rumpker and Silver, 1998; Saltzer et al., 2000).

521 The predicted delay times are smaller than the observed. Figure 9 shows a scatter-plot, comparing  
522 the predicted delay times of the surface wave model with those delay times of the smoothed,  
523 observed delay times for smoothing parameters of  $R = 600$  km. The majority of the data are  
524 located on the right-hand side of the graph, implying that surface wave-predicted delay times are  
525 biased to smaller values. This also shows up in relatively low coherence (Figure 4). Our findings  
526 of best coherence between the two datasets at approximately 600 km is in agreement with the  
527 stochastic model of anisotropy correlation by Becker et al. (2007), which reveals a relatively short  
528 correlation length of splitting parameters of  $L \sim 550$  km for orogens and volcanic zones.

529 There is only a limited amount of shear-wave splitting measurements in the oceans although future  
530 deployments of Ocean Bottom Seismometers will eventually enhance the global splitting coverage  
531 (e.g., Suetsugu et al., 2005). This should allow for a separate analysis of continents having mostly  
532 thick and old lithosphere, with sometimes ancient anisotropy orientations, and oceans having a  
533 much more recent and simpler history (Montagner, 2002; Becker et al., 2003), which will probably  
534 include exciting new possibilities such as observing hotspot-induced mantle flow, small-scale  
535 convection, plume-ridge interaction etc.

536 The coverage of shear-wave splitting data that are available at this point does not allow accepting  
537 or rejecting predominantly East-West mantle flow (e.g., Doglioni, 1994) at global scale as  
538 manifested in fast directions. The maximum observed in Figure 3b is clearly an artefact of the  
539 spatial sampling and probably includes many observations of fossil lithospheric anisotropy.

## 540 6. Conclusion

541 In this paper, we have compared shear-wave splitting measurements made around the world with  
542 predicted splitting parameters derived from anisotropic surface wave tomography. This allowed us  
543 to show that the two measures of anisotropy are indeed correlated, even though they have very  
544 different lateral and vertical resolution. This correlation is apparently supported by the fact that

545 both measures are mostly sensitive to the upper mantle (e.g., Sieminski et al., 2007). The quality of  
546 the measurements and resulting models allows us to address the limit of lateral resolution of the  
547 anisotropic surface wave model. Finally, the agreement of the two different methods highlights the  
548 fact that the two techniques are fully complementary since they are sensitive to vertical or lateral  
549 variations of upper mantle flow and therefore deserve to be twined together in studies aiming at  
550 mapping upper mantle deformation.

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558



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826 **Figure Captions**

827 Figure 1: Paths and spatial resolution associated with teleseismic (SKS) shear waves and surface  
828 waves. Surface waves propagate along the surface of the Earth, while wave paths of  
829 teleseismic shear waves are nearly vertical. For the latter, there is a progressive time delay  
830 between the two (split) shear-waves on the path toward the station at the Earth's surface.

831

832 Figure 2: Global comparison of shear-wave splitting parameters: a) measured b) predicted. a)  
833 World-wide shear-wave splitting observations contained in the splitting database. b):  
834 predicted splitting values calculated using the method of Montagner et al. (2000) for the  
835  $1^\circ \times 1^\circ$  anisotropic tomography model of Debayle et al. (2005) at depths between 50 and  
836 500km. Values are displayed on a  $3^\circ \times 3^\circ$  grid. The regions which are analysed in further  
837 detail in the text are marked (a-h)

838

839 Figure 3: Statistical distribution of 2286 globally distributed shear-wave splitting measurements  
840 contained in the current SKS splitting database. a) shows the distribution of delay times, b)  
841 that of fast split directions. The statistical distributions of predicted splitting parameters,  
842 calculated from surface wave anisotropy (Debayle et al., 2005) are presented separately  
843 for continents (c, d) and oceans (e, f). See text for discussion.

844

845 Figure 4: Coherence of the SKS splitting parameters with the predicted splitting values integrating  
846 anisotropy between 50 to 500 km depth, for “correlation lengths”  $D_{\text{corr}}$  of 5, 10, and 20  
847 degrees. Several combinations of datasets are tested, such as the effect of a  $1^\circ$  and  $10^\circ$  grid  
848 of predicted values and a limited set of data (e and f): studies published before 1997,  
849 corresponding to dataset of Silver (1996)). See text for discussion.

850

851 Figure 5: Smoothed version of shear-wave splitting observations. Smoothing parameters are  
852 chosen as  $R = 600$  km and  $\lambda = 5$  (cf. Figure 7). Calculation was for a  $1^\circ \times 1^\circ$  grid, displayed  
853 on a  $3^\circ \times 3^\circ$  grid

854

855 Figure 6: a) Coherence function using the smoothed splitting data of Figure 5. Three different  
856 values for  $R$  are used (100, 600, 1000 km), as well as a smoothing value  $\lambda = 5$ . b)  
857 Coherence value as a function of search smoothing parameter  $\lambda$  at  $0^\circ$  misfit. c) and d)  
858 show the same as a) and b) respectively, but for a randomised distribution of splitting  
859 parameters to test for inherent correlation of the two models. Note the much lower  
860 amplitude of correlation in c) and the random variation in d).

861

862 Figure 7: Correlation of the smoothed splitting parameters (similar to Figure 4), for individual  
863 regions shown in Figure 5.

864

865 Figure 8: Map of the predicted (blue), observed (red, with circle marker) and smoothed (magenta  
866 with small arrow head) shear wave splitting data in a) Western US and b) Kaapvaal Craton  
867 where the tomographic model (i.e., predicted splitting) seems to be dominated by the  
868 African Super Swell. Differences in coherence magnitude (Figure 7) are caused by  
869 different delay time due to damping effects in the tomography.

870

871 Figure 9: Scatter-plot of predicted splitting delay versus observed splitting delay. The predicted  
872 delay times are generally smaller, indicating a bias from the tomography, probably due to

873 damping effects.

874

875 Figure A1: Map of the predicted (blue), observed (red, with circle marker) and smoothed (magenta  
876 with small arrow head) shear wave splitting data in the Andes.

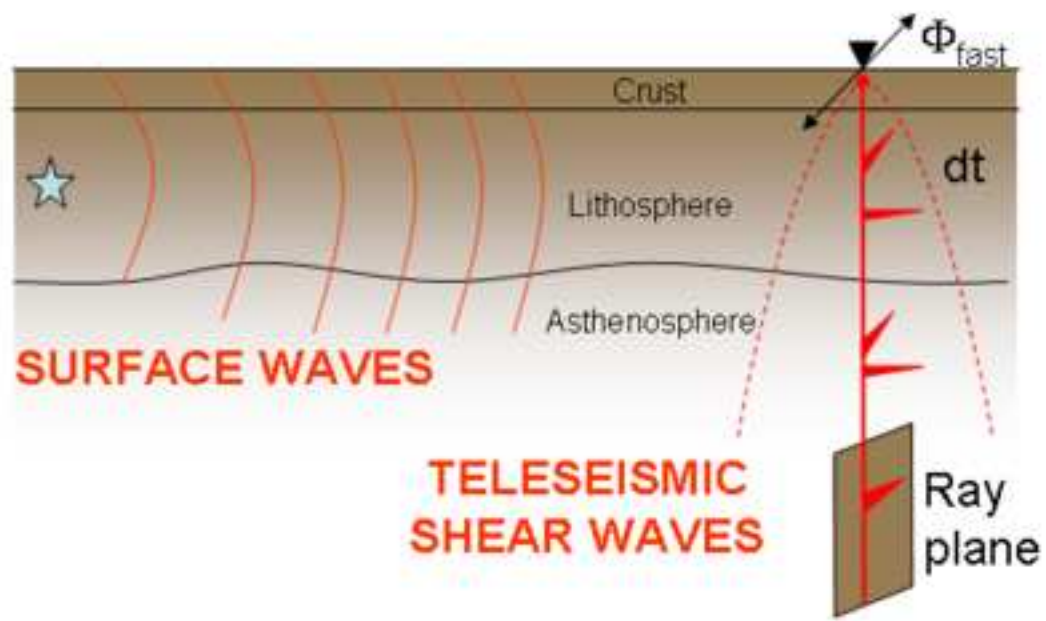
877 Figure A2: Map of the predicted (blue), observed (red, with circle marker) and smoothed (magenta  
878 with small arrow head) shear wave splitting data in the East Pacific.

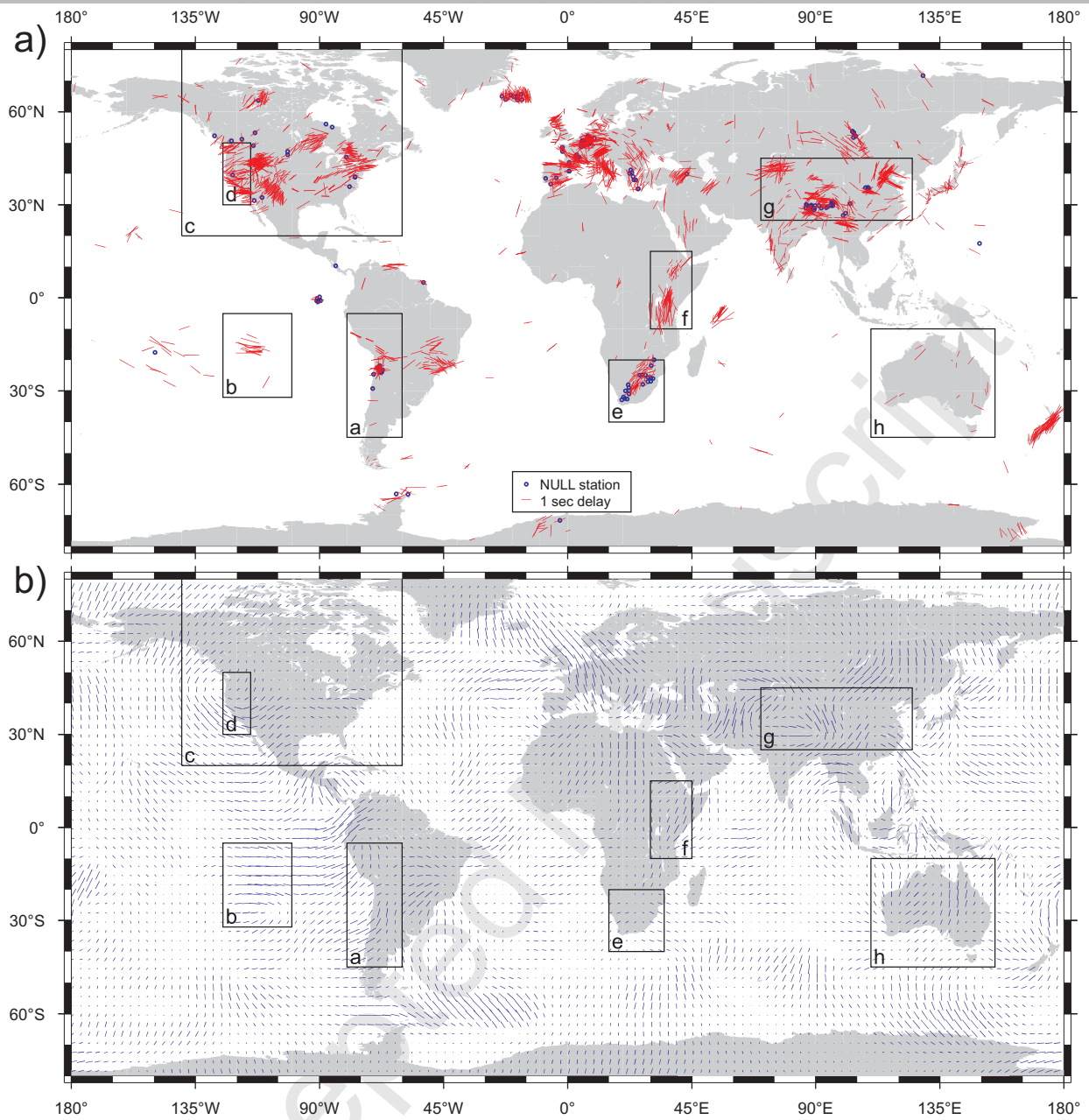
879 Figure A3: Map of the predicted (blue), observed (red, with circle marker) and smoothed (magenta  
880 with small arrow head) shear wave splitting data in the North America.

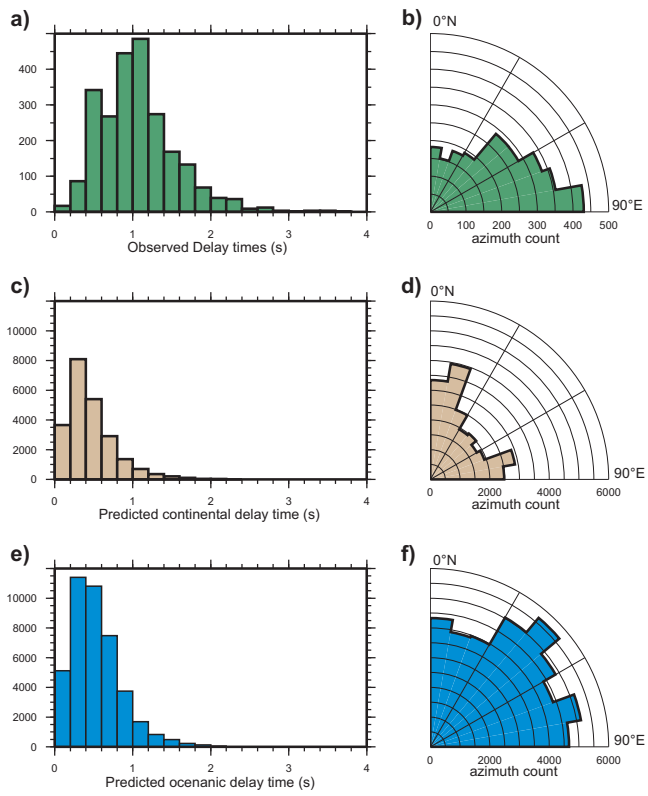
881 Figure A4: Map of the predicted (blue), observed (red, with circle marker) and smoothed (magenta  
882 with small arrow head) shear wave splitting data in East Africa.

883 Figure A5: Map of the predicted (blue), observed (red, with circle marker) and smoothed (magenta  
884 with small arrow head) shear wave splitting data in Central Asia.

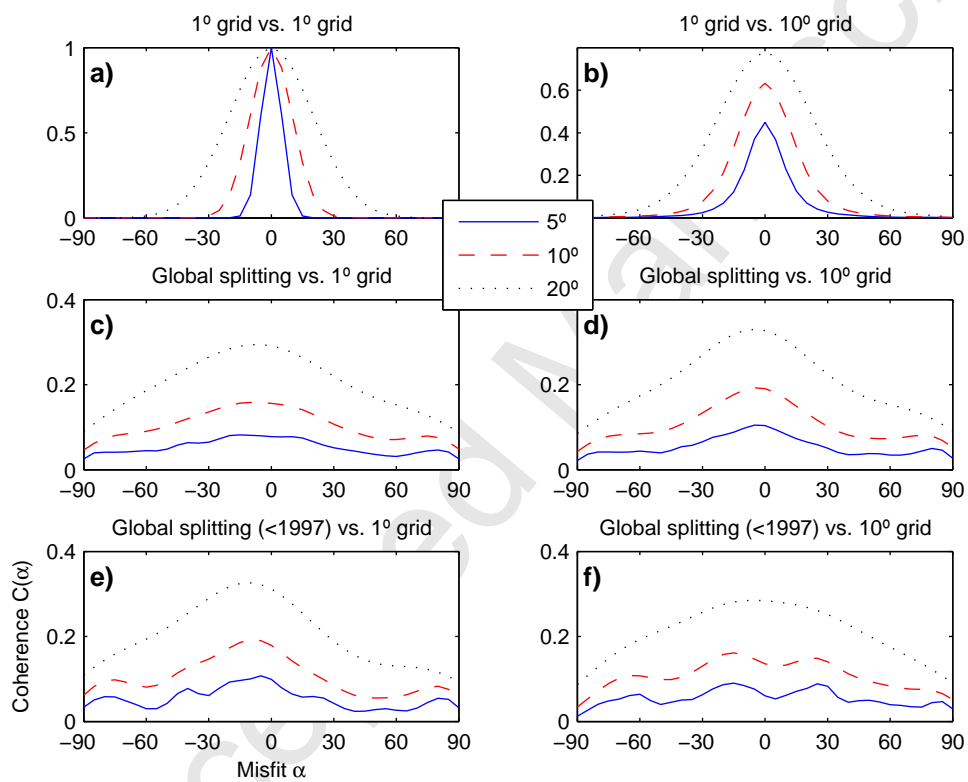
885 Figure A6: Map of the predicted (blue), observed (red, with circle marker) and smoothed (magenta  
886 with small arrow head) shear wave splitting data in Australia

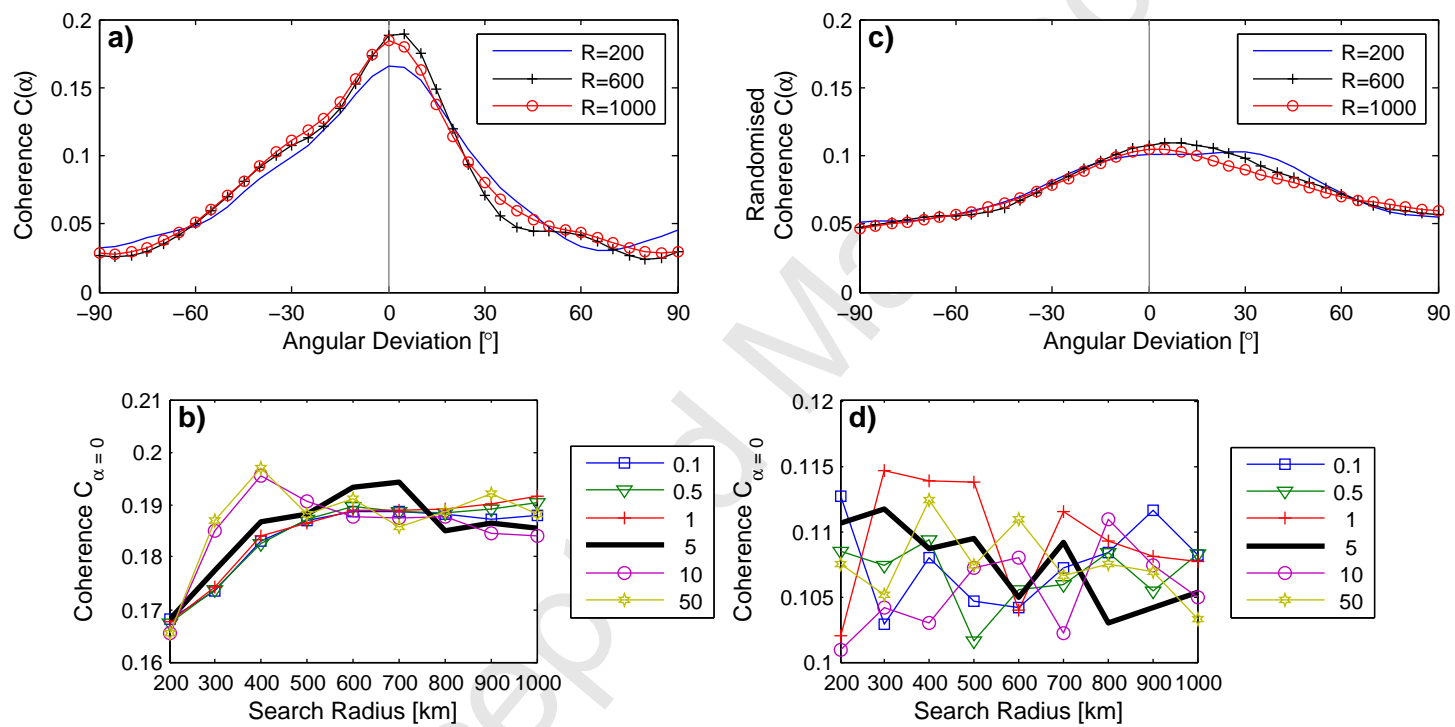


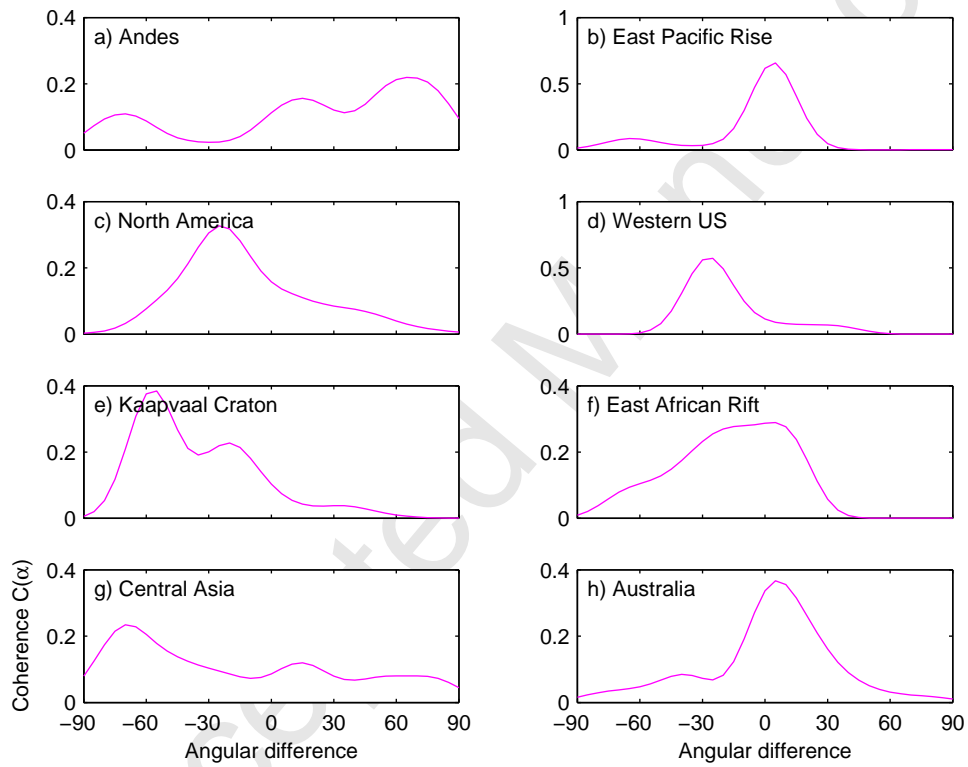


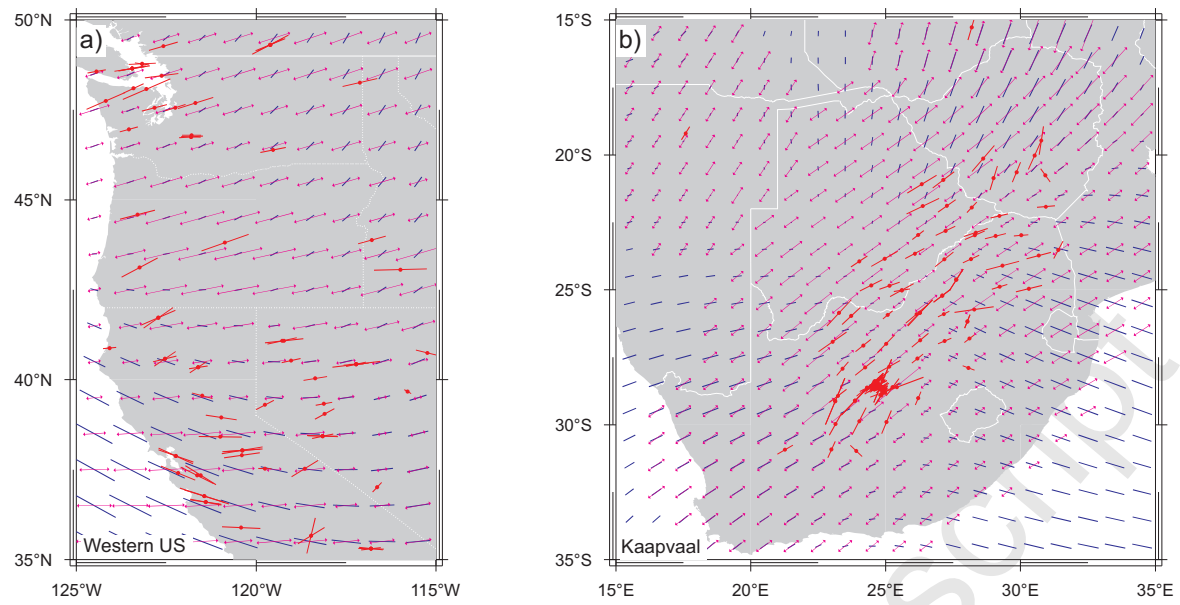


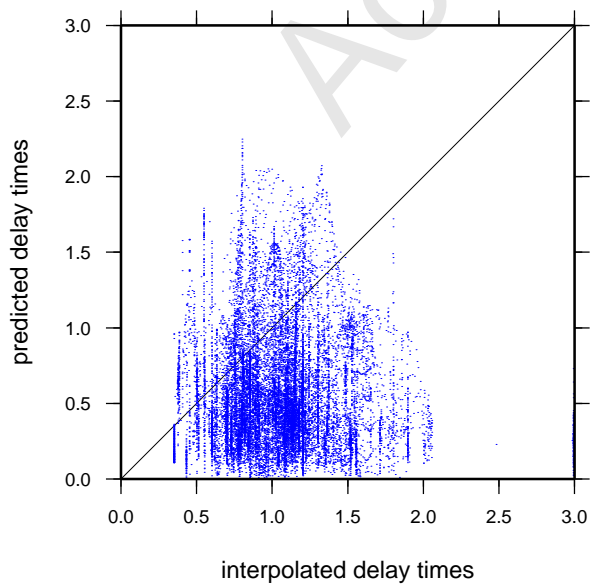


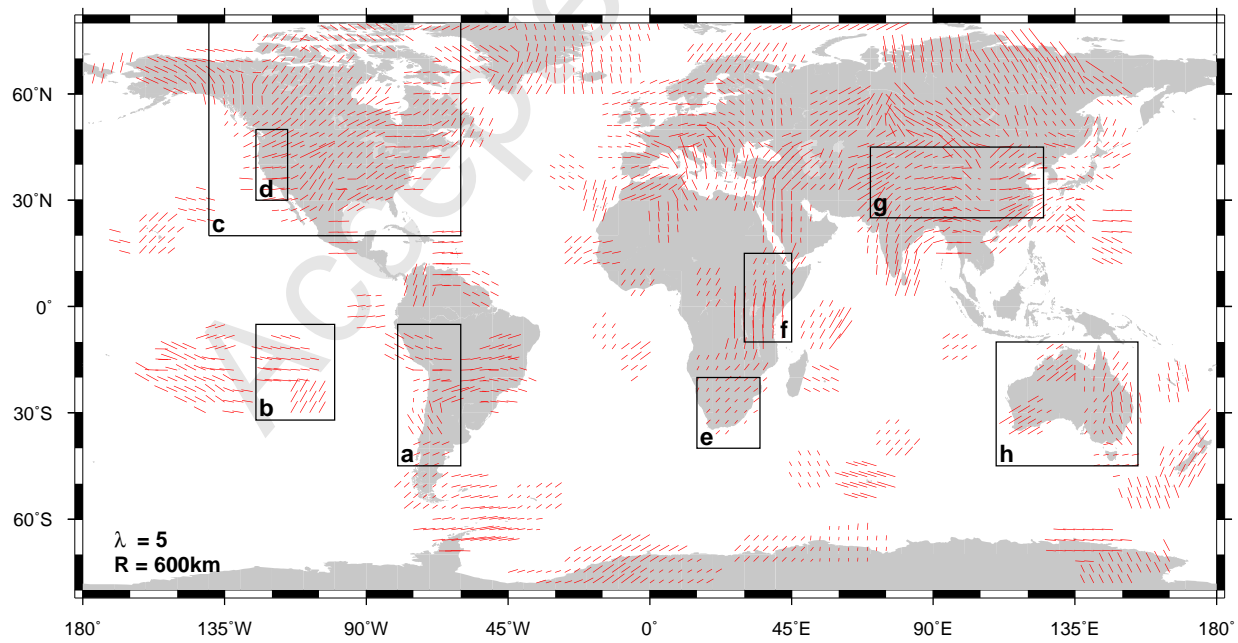


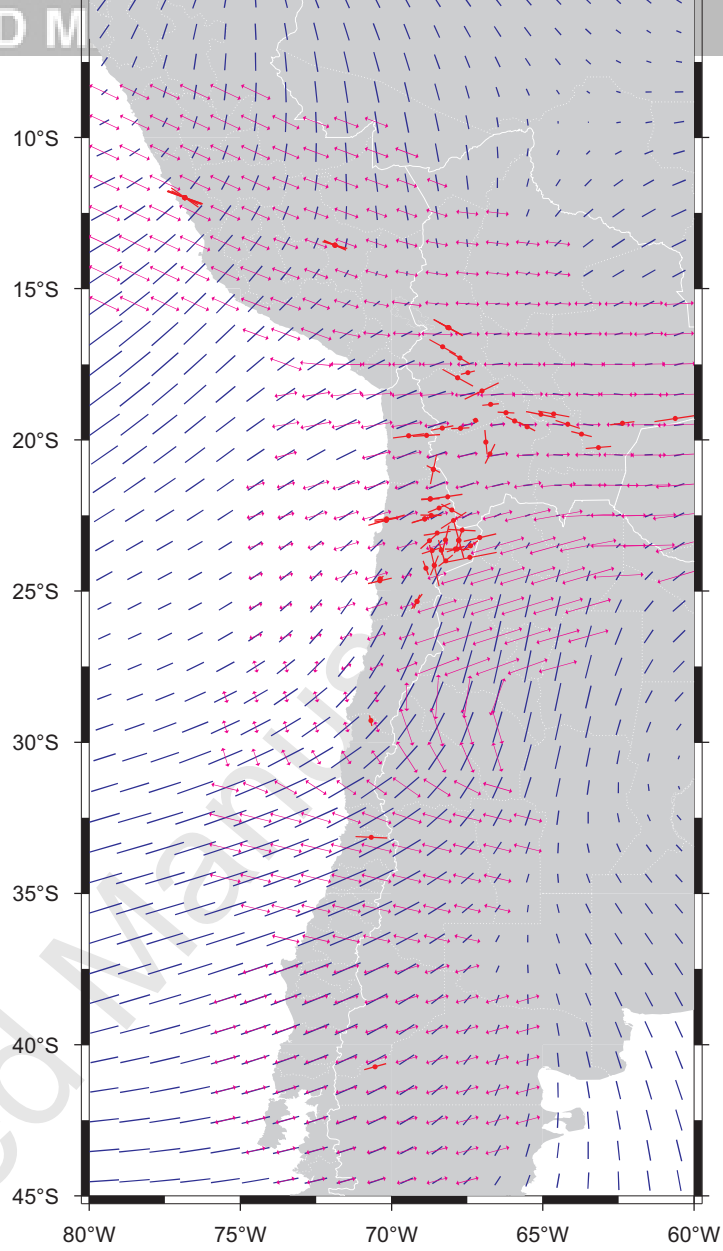


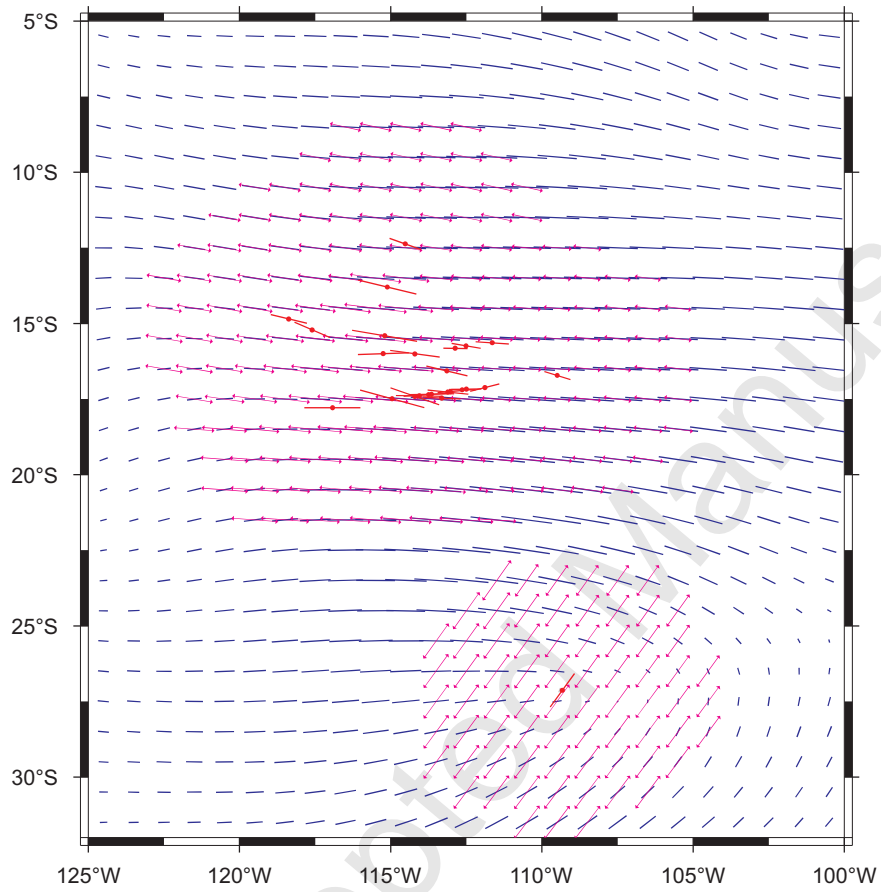




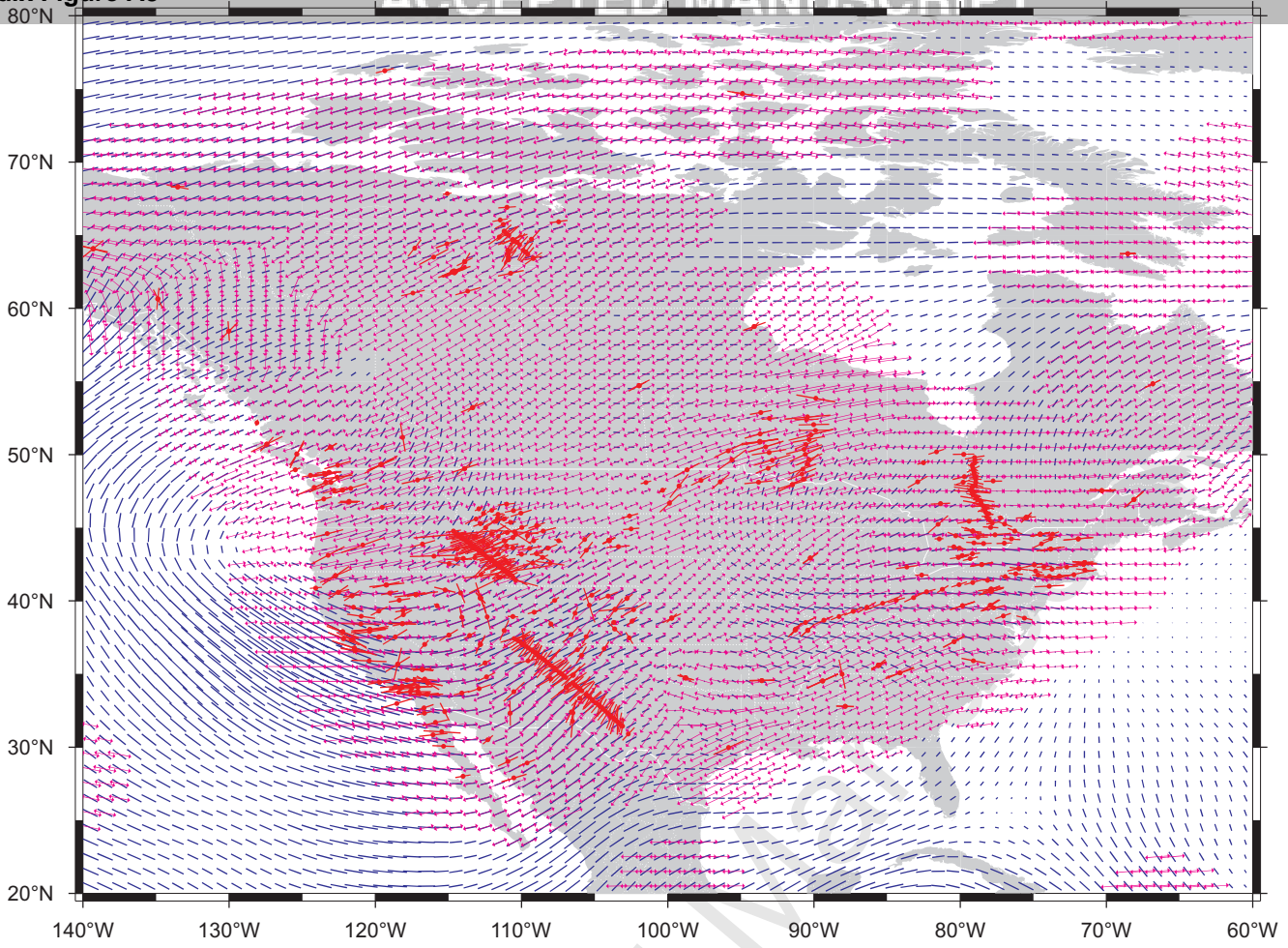




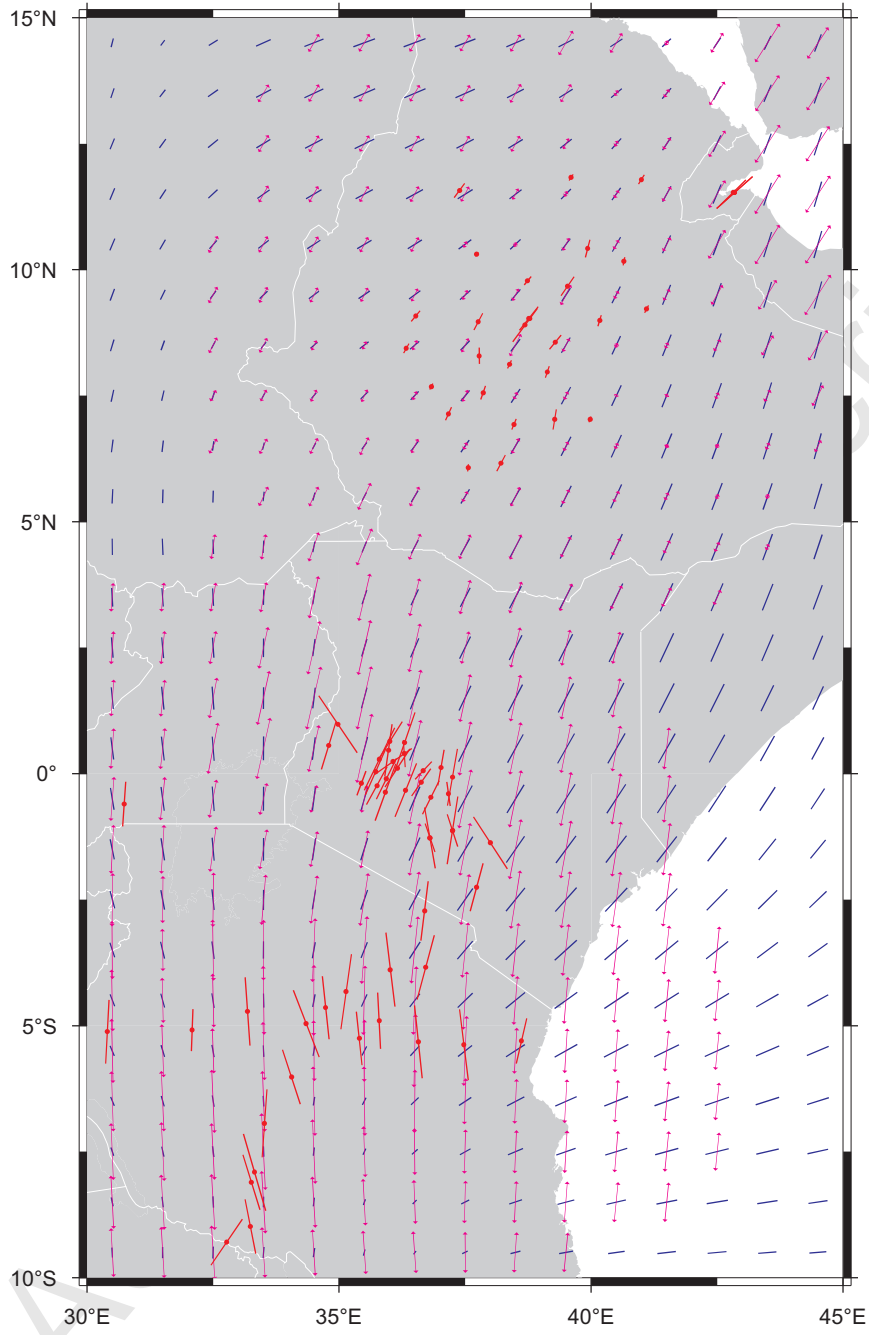


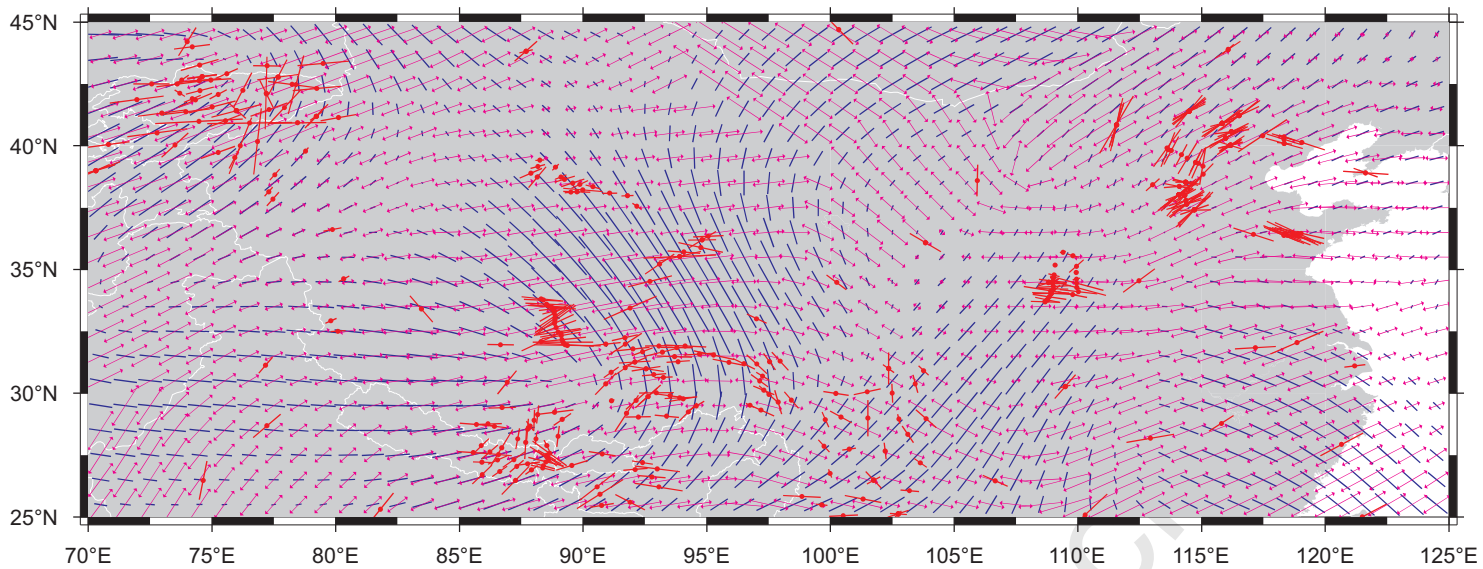






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