1 Crustal anisotropy and ductile flow beneath the eastern
2 Tibetan Plateau and adjacent areas

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9 **Abstract**

10 Crustal anisotropy beneath 71 broadband seismic stations situated at
11 the eastern Tibetan Plateau and the Sichuan Basin is investigated based on
12 the sinusoidal moveout of P-to-S conversions from the Moho and an intra-
13 crustal discontinuity. Significant crustal anisotropy is pervasively detected
14 beneath the study area with an average splitting time of 0.39 ± 0.18 s. The
15 resulting fast orientations are mostly parallel to the major shear zones in
16 the Songpan-Ganzi Terrane, and can be explained by fluid-filled fractures,
17 favoring the model of rigid block motion with deformations concentrated
18 on the block boundaries. In the vicinity of the Xianshuihe-Xiaojiang Fault
19 Zone in the southern Songpan-Ganzi Terrane, our results, when combined
20 with previously revealed high crustal Poisson's ratio in the area, support
21 the existence of mid/lower crustal flow. The Longmenshan Fault Zone and
22 adjacent areas are dominated by strike-orthogonal fast orientations, which
23 are consistent with alignments of cracks associated with compressional stress
24 between the Plateau and the Sichuan Basin. The observations suggest that

25 crustal thickening is the main cause of the high topographic relief across the
26 Longmenshan Fault Zone.

27 *Keywords:*

28 Seismic anisotropy, Crustal flow, Tibetan Plateau, Receiver functions

29 **1. Introduction**

30 The growth of the Tibetan Plateau, which has an average elevation of
31 ~ 5000 m, is widely considered to be the result of progressive collision between
32 the Indian and Eurasian plates, starting approximately 70 Ma (e.g., Yin and
33 Harrison, 2000). N-S lithospheric shortening associated with the collision has
34 resulted in eastward extrusion of the Plateau since about 15 - 20 Ma (Royden
35 et al., 2008). Global Positioning System (GPS) velocity measurements (Gan
36 et al., 2007) show eastward extrusion of central Tibet at a rate of about 15
37 - 20 mm/yr relative to the Sichuan Basin (Fig. 1). The bulk of the eastern
38 Tibetan Plateau is the Songpan-Ganzi Terrane located between the Kunlun-
39 Muztagh Suture to the north, the Jinsha River Suture to the south, and the
40 Longmenshan Fault Zone to the east (Fig. 1). Blocks of the Songpan-Ganzi
41 Terrane, including Aba, Litang, and Longmenshan, move relatively from each
42 other along the sinistral Xianshuihe-Xiaojiang and dextral Longriba faults
43 (Fig. 1), partially due to the blockage by the relatively rigid Sichuan Basin.

44 A recent study (Ji et al., 2015) suggests that the southeastward move-
45 ment of the Litang Block resulted in the transformation of the E-W oriented,
46 highly-foliated structures originally located at the Eastern Himalayan Syn-
47 taxis into NW-SE oriented tectonic fabrics. The present-day surface motion
48 of the northern Songpan-Ganzi Terrane is characterized by a northward de-

creasing rate and a slight spatial variation of direction, indicating the presence of both compression and strike-slip shear deformations, which are also indicated by the pervasive thrust and strike-slip faults (Fig. 1). The eastern margin of the Tibetan Plateau, the Longmenshan Fault Zone, is characterized by a significant topographic relief relative to the plateau interior and the Sichuan Basin, with a low strike-orthogonal shortening rate of less than 3 mm/yr, but a high compressional stress rate (Gan et al., 2007). As discussed below, the fault zones contribute significantly to the observed crustal anisotropy, together with lower crustal flow and the aforementioned tectonic fabrics developed from foliated structures.

Mechanisms responsible for the expansion and uplifting of the eastern Tibetan Plateau remain enigmatic, in spite of numerous studies (e.g., Molnar and Tapponnier, 1975; England and Houseman, 1986; Royden et al., 1997; Gan et al., 2007; Bai et al., 2010; Q. Liu et al., 2014). Proposed geodynamic models applicable to eastern Tibet can be loosely divided into three groups. Those in the first group suggest a dominant role of a channeled ductile flow system in the mid/lower crust in the uplifting and lateral expansion of the Plateau (e.g., Bird, 1991; Royden et al., 1997; Clark and Royden, 2000; Burchfiel et al., 2008; Zhao et al., 2012). Those models are consistent with the general absence of upper-crustal thrust faults in the area. The second group of models advocate broadly distributed deformation of crustal shortening and thickening as a predominant mechanism (England and Houseman, 1986; Wang et al., 2008; Hubbard and Shaw, 2009). Those in the third group attribute the uplifting to successive developments of crustal thrust-wedges associated with the oblique subduction of the lithospheric mantle along ma-

74 jor fault zones (Tapponnier et al., 1982; 2001; Replumaz and Tapponnier,
75 2003). One of the key constraints on the various models is a high-resolution
76 quantification of crustal deformation, which can be characterized by seismic
77 anisotropy measurements.

78 *1.1. Formation mechanisms of crustal and mantle anisotropy*

79 Numerous studies demonstrate that seismic azimuthal anisotropy, as quan-
80 tified by the polarization orientation of the fast shear wave (ϕ) and the split-
81 ting delay time (δt) between the fast and slow shear waves, is a nearly ubiq-
82 uitous property of the Earth’s crust and mantle (Silver, 1996; Savage, 1999).
83 It is generally believed that azimuthal anisotropy in the continental upper
84 crust is mostly the result of shape preferred orientation of fluid-saturated
85 vertical cracks (Crampin, 1981), which are sub-parallel to the maximum hor-
86 izontal compression direction. The resulting δt is normally smaller than 0.2
87 s (Crampin, 1994), but can reach 0.5 s or greater for stations located in or
88 close to a fault zone (e.g., Savage et al., 1990; Liu and Niu, 2012). Highly-
89 deformed fabrics with a vertical foliation plane formed by compressional fold-
90 ing are also found to produce observable azimuthal anisotropy originating
91 from anisotropic minerals in the crust such as mica and amphibole (Ji et al.,
92 2015).

93 In the mid/lower crust, due to the closure of cracks, azimuthal anisotropy
94 is mostly produced by the lattice preferred orientation (LPO) of anisotropic
95 crystals, primarily amphibole (Tatham et al., 2008; Ko and Jung, 2015). A
96 recent study of amphibole (Ko and Jung, 2015) suggests that channelized
97 plastic flow in the mid/lower crust can result in azimuthal anisotropy. The
98 relationship between the resulting ϕ and flow direction is dependent on the

99 differential stress level and temperature (Fig. 2). Under the condition of
100 high differential stress and high temperature, which is possible for the Ti-
101 betan Plateau due to significant crustal shortening and thickening (Yin and
102 Harrison, 2000), type *II* and *III* fabrics, for which ϕ is sub-parallel to the
103 flow direction, will dominate (Fig. 2).

104 Seismic anisotropy in the mantle is generally regarded as the consequence
105 of LPO of anisotropic minerals, primarily olivine (Zhang and Karato, 1995).
106 Mantle flow results in a fast orientation that is sub-parallel to the flow di-
107 rection, and for areas experienced lithospheric shortening, the resulting ϕ is
108 frequently observed in accordance with the strike of the mountain belt (e.g.,
109 McNamara et al., 1994; Silver, 1996; Bokelmann et al., 2013).

110 1.2. Previous seismic anisotropy investigations in the study area

111 Crustal and mantle azimuthal anisotropy beneath the eastern Tibetan
112 Plateau has been investigated by a number of studies (e.g., Sol et al., 2007;
113 Wang et al., 2008; Sun et al., 2012; Chen et al., 2013; Sun et al., 2015).
114 Studies using the splitting of XKS phases (including SKS, SKKS, and PKS),
115 which measure the integrated anisotropy of the crust and mantle, reveal an
116 average splitting time of about 1.0 s. The fast orientations from XKS are
117 mostly parallel to the strike of surface geological features (Fig. 3).

118 Crustal anisotropy studies (e.g., Sun et al., 2012; Sun et al., 2015) suggest
119 that the crust at some sites can produce large splitting times comparable
120 to the global average of 1.0 s from XKS studies, suggesting a significant
121 crustal contribution to the observed XKS splitting. A crustal anisotropy
122 investigation (Sun et al., 2015) in the vicinity of the Longmenshan Fault
123 Zone at 21 stations using the P-to-S converted phases from the Moho reports

124 significant crustal anisotropy, with δt values ranging from 0.22 to 0.94 s. The
125 observed anisotropy is attributed to the LPO of mica or amphibole deformed
126 by lower crustal flow. The same formation mechanism of crustal anisotropy
127 has been suggested beneath the Xianshuihe-Xiaojiang Fault Zone (e.g., Sun
128 et al., 2012; Chen et al., 2013). In contrast, Ji et al. (2015) proposes that sub-
129 vertical foliation planes, which lead to transverse isotropy with a horizontal
130 axis of symmetry, are responsible for the observed crustal anisotropy beneath
131 the study area.

132 In this study, crustal azimuthal anisotropy beneath the eastern Tibetan
133 Plateau and adjacent areas is explored based on the sinusoidal moveout of
134 the P-to-S converted phases from the Moho and (at a few stations) an intra-
135 crustal discontinuity (Rumpker et al., 2014). The measurements have an
136 average δt of 0.39 ± 0.18 s, and show dominantly fracture-parallel fast orien-
137 tations.

138 **2. Data and methods**

139 *2.1. Data*

140 The broadband seismic data used in the study were recorded by 71 sta-
141 tions (Fig. 1), of which 53 were provided by the Data Management Centre
142 of China National Seismic Network at Institute of Geophysics, China Earth-
143 quake Administration for the recording period of 2007-2011. Data from the
144 other 18 stations were obtained from the Incorporated Research Institutions
145 for Seismology (IRIS) Data Management Center (DMC) for the period of
146 2003-2014. All the teleseismic events with magnitude ≥ 4.0 and epicentral
147 distance in the range of 30° - 180° are band-pass filtered in the frequency

band of 0.08 - 0.8 Hz, and those with visible first P-arrivals on the vertical component are converted to radial receiver functions (Ammon, 1991), which are visually inspected to select the ones with clear P-arrivals and without anomalously large arrivals in the P-wave coda. The two phases utilized in this study include P_ms , which is the P-to-S converted phase from the Moho for measuring the bulk crustal anisotropy, and P_is , which is the P-to-S converted phase from an intra-crustal interface for detecting upper crustal anisotropy at a few stations where P_is is clearly observed.

2.2. Detection of single layer of anisotropy

For a single layer of anisotropy with a horizontal axis of symmetry, the arrival times of the P-to-S converted phase (P_ms or P_is) vary systematically with the back-azimuth (BAZ) of the events (Liu and Niu, 2012; Rumpker et al., 2014), i.e.,

$$t = t_0 + \Delta t = t_0 - \frac{\delta t}{2} \cos[2(\alpha - \phi)], \quad (1)$$

where t_0 represents the arrival time in the isotropic case, Δt is the offset caused by crustal anisotropy along the raypath, δt is the delay time which reflects the strength of crustal anisotropy, ϕ is the fast orientation of crustal anisotropy measured clockwise from the north, and α is the BAZ of the events.

In this study, two pre-processing steps are applied to enhance the reliability of the measurements. First, in order to eliminate the moveout caused by epicentral distances, all the receiver function traces are adjusted to a uniform epicentral distance of 60°. Second, traces belonging to the same BAZ

170 window of 10° wide are stacked together to enhance the signal-to-noise ratio.
 171 The splitting parameters, ϕ and δt , can be obtained by fitting the P_ms (or
 172 P_is) arrival times relative to the direct P-wave using Equation (1) based on a
 173 non-linear least-squares fitting procedure. An example can be found in Fig.
 174 4.

175 To quantify the uncertainties of the resulting crustal anisotropy measure-
 176 ments, we apply the bootstrap resampling procedure (Efron and Tibshirani,
 177 1986; Press et al., 1992) to estimate the splitting parameters 10 times. The
 178 standard deviation (SD), σ , for the station is quantified by a unit-less value
 179 computed using

$$\sigma = \frac{\sigma_\phi}{90.0} + \frac{\sigma_{\delta t}}{1.0}, \quad (2)$$

180 where σ_ϕ and $\sigma_{\delta t}$ are the SD of ϕ and δt , respectively, estimated using the
 181 bootstrap procedure. Only results from stations with a $\sigma \leq 0.4$ are presented
 182 and discussed below.

183 *2.3. Characterization of two layers of anisotropy*

184 Two layers of crustal anisotropy are detected and characterized at 3 of the
 185 stations where a P_is phase is clearly identified in addition to the P_ms (Fig.
 186 5), following the layer-stripping technique of Rumpker et al. (2014). For a
 187 given station with both P_ms and P_is arrivals, the P_is phases are used first to
 188 constrain the anisotropy in the layer above the interface ("upper crust"). The
 189 P_ms arrival times, which reflect the integrated effects of anisotropy in both
 190 the upper and lower crust, are then corrected using the resulting anisotropy
 191 parameters of the upper crust to remove the contributions of upper crustal

192 anisotropy. The anisotropy of the lower layer can then be quantified by
193 applying Equation (1) to the corrected P_m s travel-times (Fig. 5).

194 In order to derive robust two layer crustal anisotropy measurements, some
195 preconditions need to be satisfied: (1) the existence of an intra-crustal dis-
196 continuity, (2) clear and azimuthally varying P_i s and P_m s arrivals, and (3)
197 good back-azimuthal coverage of high-quality radial receiver functions. Only
198 three of the stations, CD2, MAQU, and MEK, which are all located in the
199 vicinity of known fault zones (Fig. 3), lead to reliable characterization of
200 two-layer crustal anisotropy structure.

201 3. Results

202 Well-defined crustal anisotropy is measured at a total of 71 stations (Fig.
203 3). While the resulting ϕ measurements are dominantly parallel to the strike
204 of major surface features, a sharp contrast of the δt values, which have an
205 average value 0.39 ± 0.18 s, is observed between the Tibetan Plateau and the
206 Sichuan Basin separated by the Longmenshan Fault Zone, with an mean
207 value of 0.44 ± 0.18 s for the west, and 0.27 ± 0.13 s for the east (Fig. 6).

208 Comparison between our results and those by Chen et al. (2013) and Sun
209 et al. (2015) are shown in Fig. 7. Chen et al. (2013) provided 33 crustal
210 anisotropy measurements in our study area, and Sun et al. (2015) reported 21
211 measurements along the Longmenshan Fault Zone. While some differences
212 exist at some stations (Fig. 7), the measurements from the 3 studies are
213 generally consistent, in spite of the significantly larger spatial coverage that
214 this study provides. As detailed in the Discussion section, the increased
215 spatial coverage is essential for constraining anisotropy-forming mechanisms,

216 as well as for providing valuable constraints for crustal deformation models.

217 Based on the major geologic features, we divide the study area into four
218 regions, including the Kunlun-Muztagh Suture Zone, Xianshuihe-Xiaojiang
219 Fault Zone, Longmenshan Block and adjacent areas, and the Sichuan Basin
220 (Fig. 1).

221 *3.1. Kunlun-Muztagh Suture Zone and Xianshuihe-Xiaojiang Fault Zone*

222 Crustal anisotropy measurements from the 28 stations situated in the
223 Kunlun-Muztagh Suture Zone (Fig. 1) result in an average δt of 0.47 ± 0.18
224 s. The ϕ measurements are dominantly ESE-WNW, consistent with the
225 strike of the major faults. Fault-parallel fast orientations are also obtained at
226 several stations in the same area by a recent crustal anisotropy study (Wang
227 et al., 2016). Station MAQU demonstrates the existence of two layers of
228 anisotropy (Fig. 5), with comparable ϕ and δt values between the 2 layers.
229 The similarity of anisotropy between the upper and lower crust may suggest
230 that beneath the station, the entire crust deforms coherently.

231 Similarly, the Xianshuihe-Xiaojiang Fault Zone in southern Songpan-
232 Ganzi Terrane also exhibits large δt measurements (0.53 ± 0.16 s) and strike-
233 parallel fast orientations. The Kunlun-Muztagh Suture Zone and Xianshuihe-
234 Xiaojiang Fault Zone possess the largest δt values in the study area (Fig. 8).

235 *3.2. Longmenshan Block and adjacent areas*

236 The Longmenshan Block and adjacent areas between the Kunlun-Muztagh
237 Suture Zone and the Xianshuihe-Xiaojiang Fault Zone are sampled by 20 sta-
238 tions. The ϕ measurements are dominantly NW-SE and the rest are mostly
239 consistent with the NE-SW strike of faults (Fig. 3). The measurements from

the 3 stations located at the Longriba Fault Zone are characterized by small δt with a mean value of 0.17 ± 0.03 s, and the fast orientations are either parallel or orthogonal to the strike of the faults. Measurements at station MEK (Fig. 5) suggest the presence of two anisotropic layers with orthogonal ϕ and comparable δt measurements. The ϕ of the lower layer is normal to the strike direction of the fault zone, while it is strike-parallel for the upper layer.

The ϕ measurements from the 3 stations situated in the Longmenshan Fault Zone are parallel to the NE-SW strike of the faults, while the 8 stations situated in the adjacent areas of the faults are strike-orthogonal. The ϕ of the upper layer anisotropy beneath station CD2 is parallel to the strike of the Longmenshan Fault Zone, and that of the lower layer is orthogonal to it. Generally, the δt values along the Longmenshan Fault Zone are smaller than those obtained in the Kunlun-Muztagh Suture Zone and Xianshuihe-Xiaojiang Fault Zone, and the average δt is 0.35 ± 0.13 s (Fig. 8).

3.3. Sichuan Basin

The 15 stations located in the Sichuan Basin lead to a mean δt value of 0.25 ± 0.13 s, with smaller δt (≤ 0.1 s) at stations in the interior of the basin. The ϕ measurements are mostly parallel to the surface fabric. The results are similar to those obtained by Sun et al. (2015) at 3 stations in the Sichuan Basin.

261 4. Discussion

262 4.1. Formation mechanisms of the observed crustal anisotropy and geody- 263 namic implications

264 As discussed previously, a number of processes can lead to observable
265 crustal anisotropy, including fluid-filled fracture zones, vertical foliation planes
266 containing anisotropic minerals, and mid/lower crustal flow that aligns anisotropic
267 minerals. For a given area, the relative importance of the processes is de-
268 pendent on the tectonic setting and crustal properties such as thickness and
269 Vp/Vs ratio (γ) which is uniquely related to the better-known Poisson's ratio
270 (σ) by $\sigma = 0.5[1 - 1/(\gamma^2 - 1)]$. In this section, we speculate on the formation
271 mechanisms for each of the areas on the basis of existing surface geology and
272 crustal property observations, as well as their relationship with the observed
273 crustal anisotropy parameters.

274 4.1.1. Kunlun-Muztagh Suture Zone

275 Significant crustal anisotropy is observed throughout the Kunlun-Muztagh
276 Suture Zone, with the ϕ measurements being dominantly parallel to the ma-
277 jor faults in the suture zone. Previous receiver function studies of the crust
278 report Vp/Vs values of about 1.73 which are lower than the global average
279 of 1.76 for continental crust (e.g., Wang et al., 2010; Sun et al., 2015), sug-
280 gesting the absence of large-scale crustal partial melting, which would lead
281 to much higher Vp/Vs values (e.g., Reed et al., 2014). Vertically coherent
282 crustal-mantle deformation and a possible absence of lower crustal flow be-
283 neath this area are also suggested by a recent study of Wang et al. (2016),
284 who observed low crustal Vp/Vs ratio and strong fault-parallel crustal and

285 mantle anisotropy.

286 To explore the relationship between crustal anisotropy measurements and
287 the faults, we plot the observed splitting times against the distance of the
288 stations from the nearest fault. As shown in Fig. 6b, for stations within 20
289 km from the nearest fault, the splitting times are the largest (about 0.6 s)
290 and are not varying with the distance, implying the existence of a broad shear
291 zone at depth. For stations that are more than 20 km away from the fault,
292 there seems to be a gradual decrease in the splitting times with increasing
293 distance. This relationship, together with the fact that the fast orientations
294 are mostly parallel to the strike of the faults in the suture zone, suggests that
295 crustal anisotropy in this area is mostly related to fractures associated with
296 the suture zone.

297 4.1.2. *Xianshuihe-Xiaojiang Fault Zone*

298 Similar to the Kunlun-Muztagh Suture Zone, the Xianshuihe-Xiaojiang
299 Fault Zone is characterized by fault-strike-parallel fast orientations (Fig. 3)
300 and large splitting times (Fig. 8). However, unlike the former, an inverse
301 relationship between the splitting times and the distance to the nearest fault
302 is not robustly observed. This could indicate that unlike those in the suture
303 zone, fractures in the Xianshuihe-Xiaojiang Fault Zone have a limited lateral
304 extent and thus contribute insignificantly to the observed crustal anisotropy.

305 Another difference between the two areas is that while the former has
306 lower than normal crustal V_p/V_s values, the latter is dominated by high
307 V_p/V_s ratio (≥ 1.78 ; Fig. 3) and thus may indicate the existence of crustal
308 partial melting (e.g., Wang et al., 2010; Sun et al., 2015). This is indepen-
309 dently evidenced by the high electrical conductivity observed beneath this

310 area (Bai et al., 2010), and may reflect plastic flow in the mid/lower crust
311 toward the southeast (Clark and Royden, 2000; Bai et al., 2010). The consis-
312 tency between the fast orientations and the predicted flow direction suggests
313 that the dominant LPO of amphibole in the area is Type *II* or Type *III* (Fig.
314 2), which results in a flow-parallel fast orientation for a vertically propagating
315 S-wave (Ko and Jung, 2015).

316 Besides faults in the upper crust and plastic flow in the mid/lower crust,
317 another potential contributor to the observed anisotropy in this area is shear-
318 related mineral lineation. A recent study (Ji et al., 2015) reveals sub-
319 vertically aligned foliation planes with nearly horizontal lineations in the
320 southeastern Tibetan Plateau, which are being rotated through strike-slip
321 shear around the Himalayan Syntaxis. The NW-SE oriented foliation planes
322 contain mica- and amphibole-bearing metamorphic rocks, leading to signif-
323 icant transverse isotropy with a horizontal axis of symmetry. The resulting
324 fast orientation for a vertically propagating S-wave is in the foliation plane
325 and parallel to the lineation direction.

326 4.1.3. Longmenshan Block and adjacent areas

327 Relative to other areas of the eastern Tibetan Plateau, this area has the
328 weakest crustal anisotropy (Fig. 8), in spite of the fact that it is experiencing
329 a high rate of NW-SE directed compressional stress (Gan et al., 2007). The
330 consistency between the fast orientations and the direction of the maximum
331 horizontal compression suggests that NW-SE oriented extensional cracks in
332 the upper crust are mostly responsible for the observed anisotropy. Such a
333 mechanism is commonly involved to explain crustal anisotropy in areas dom-
334 inated by compressional tectonics (Crampin, 1981; 1994). Station LORI in

335 the Longriba Fault Zone and a few stations along the Longmenshan Fault
336 show fault-parallel fast orientations and small splitting times, probably re-
337 lated to fluid-filled fractures in the fault zones. This mechanism can also
338 explain the NE-SW directed fast orientation of the upper layer at stations
339 MEK and CD2. Alternatively, the 90° difference between upper and lower
340 crustal anisotropy can be caused by a previously revealed 90° flip as a re-
341 sult of pore pressure exceeding the maximum horizontal compressional stress
342 (Crampin et al., 2003). This mechanism implies a high pore pressure in the
343 upper crust at the stations with two-layer anisotropy.

344 Crustal anisotropy studies (Shi et al., 2009; Gao et al., 2014) using shear
345 waves from local earthquakes show mostly fault-parallel fast orientations at
346 stations located in the north-most 100 km of the Longmenshan Fault zone,
347 while our results are mostly fault-orthogonal in the same area. This appar-
348 ent discrepancy could be caused by the fact that the former studies measure
349 the anisotropy of the seismogenic upper crust, while our results are inte-
350 grated over the whole crust. The fact that the crustal anisotropy has a
351 fault-orthogonal fast orientation suggests that the lower crust has a NW-SE
352 oriented anisotropy with a much larger splitting time than the upper crust.

353 The small splitting times may suggest the absence of strong mid/lower
354 crustal flow beneath this area. The blockage of the Sichuan Basin might be
355 responsible for the absence, although the study alone cannot rule out the
356 existence of partial melting in the mid/lower crust. However, the normal
357 V_p/V_s values observed in most of the area (Wang et al., 2010) place doubts
358 on the existence of significant crustal partial melting (e.g., Clark and Royden,
359 2000; Zhao et al., 2012), but support a crustal thickening model for the

360 dramatic topographic relief and recent large earthquakes in the Longmenshan
361 Fault Zone (e.g., Hubbard and Shaw, 2009; Wang et al., 2012).

362 4.1.4. *Sichuan Basin*

363 The crust beneath the Sichuan Basin is weakly anisotropic, as indicated
364 by the smallest average splitting time in the entire study area (Fig. 8), an ob-
365 servation that is consistent with the previously suggested strong lithosphere
366 beneath the basin (Huang et al., 2015). Most of the fast orientations in the
367 basin are NE-SW which might reflect the strike of crustal fabrics formed
368 by ancient tectonic events. The several stations near the SW edge of the
369 basin show edge-parallel fast orientations and could indicate the existence
370 and reflect the effects of major boundary faults.

371 4.2. *Implications on mantle deformation*

372 For all of the study area, with possible exception of the Sichuan Basin, the
373 fast orientation of crustal anisotropy is dominantly consistent with that of the
374 XKS measurements, which represent the integrated anisotropy of both the
375 crust and mantle. As shown in Fig. 3, in which the $P_m s$ and XKS splitting
376 parameters are plotted using the same scale, although crustal anisotropy is
377 a significant contributor to the observed XKS anisotropy for most areas, the
378 fact that XKS splitting times are constantly greater than $P_m s$ splitting times
379 suggests that the mantle is also anisotropic.

380 The observed mantle anisotropy can be explained by two models. The
381 first is the coherent lithospheric deformation model for which both the crustal
382 and mantle portions of the lithosphere deform as an entity (Silver, 1996;
383 Wang et al., 2008). For the mantle portion, this model attributes the suture-

384 or fault-parallel mantle anisotropy to strike-orthogonal compression, which
385 orients the olivine a-axis to the strike direction (Silver, 1996). Because the
386 dominant faults are strike-slip, and strike-slip faults are inefficient in re-
387 orienting mantle anisotropic minerals (e.g., K. Liu et al., 2014 for the San
388 Andreas Fault), it is difficult to attribute the observed mantle anisotropy to
389 coherent lithospheric deformation.

390 The second and our preferred model is that mantle anisotropy is caused
391 by simple shear in the upper-most asthenosphere due to its differential move-
392 ment with the lithosphere. Such a mechanism is widely used to explain XKS
393 anisotropy (e.g., Refayee et al., 2014 for central North America, and Lemnifi
394 et al., 2015 for North Africa). The differential movement can be induced by
395 lithospheric movement along the strike-slip faults over a stationary astheno-
396 sphere, or by a faster- or slower-moving asthenospheric flow in the direction
397 of the major fault zones.

398 5. Conclusions

399 P-to-S conversion from the Moho and an intra-crustal discontinuity re-
400 veals strong and spatially varying crustal anisotropy beneath the eastern
401 Tibetan Plateau and Sichuan Basin, with unprecedented spatial resolution.
402 For most of the study area, the resulting fast orientations are dominantly par-
403 allel to the strike of major strike-slip faults, and the largest splitting times
404 are found in major shear zones. Beneath the vicinity of the Xianshuihe-
405 Xiaojiang Fault Zone, our results support the existence of mid/lower crustal
406 flow. The pervasive strike-orthogonal fast orientations in the vicinity of the
407 Longmenshan Fault Zone suggest a significant contribution of fault-normal

408 compressional stress to the large topographic relief across the fault.

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Figure 1: A topographic relief map of the study area showing the suture zones (dashed purple lines), major faults (solid black lines), and seismic stations (triangles) used in the study. Different colors are assigned based on the sub-area that the station belongs to. Red: Kunlun-Muztagh Suture Zone; Green: Longmenshan Block and adjacent areas; Blue: Xianshuihe-Xiaojiang Fault Zone; Orange: Sichuan Basin. Light blue arrows show GPS velocities (Gan et al., 2007) relative to the solid blue circle in the Sichuan Basin. The Longriba Fault is digitized based on Ren et al. (2013), and the other faults are based on Styron et al. (2010). Dashed line in the inset map shows the location of the study area within Eurasia.

Figure 2: (a) Types of amphibole fabrics as a function of differential stress and temperature. (b) Predicted seismic anisotropy formed by a horizontal flow system for a vertically propagating S-wave (Ko and Jung, 2015). On the top plane, the fast orientation is shown in red, and the flow direction is parallel to the gray bar.

Figure 3: Crustal anisotropy measurements (red) with orientation of the bars representing the ϕ , and the length proportional to the δt values. At stations MEK and CD2, the blue bar shows anisotropy of the upper layer, and the purple bar shows that of the lower layer. The gray bars are the station-averaged XKS splitting results obtained from <http://splitting.gm.univ-montp2.fr/DB/public/searchdatabase.html>. The area southwest of the green line is dominated by high V_p/V_s ratio (≥ 1.78) from Wang et al. (2010). The orange dashed arrows show the location of proposed flow channel from Bai et al. (2010). The dashed line traversing the Longmenshan Fault Zone separates areas with large and small splitting times.

Figure 4: Receiver functions recorded by station LTT. The black dots show the peak locations of the P_m s and the red line is the theoretical P_m s moveout calculated based on Equation (1) using the optimal pair of splitting parameters.

Figure 5: Two-layer crustal anisotropy measurements using data recorded by stations MAQU (a-d) and MEK (e-h). (a) and (e) are the original receiver functions showing both the P_m s and P_i s phases; (b) and (f) show the P_i s arrivals and the fitting curve; (c) and (g) are the original P_m s arrivals, and (d) and (h) are P_m s arrivals after correcting for upper crustal anisotropy.

Figure 6: (a) Splitting times from all stations with measurements plotted against the distance from the Longmenshan Fault Zone (the dashed line in Fig. 3). Thick horizontal bars represent mean values, and thin bars represent $\text{mean} \pm \text{SD}$. (b) Splitting times from stations situated in the Kunlun-Muztagh Suture Zone plotted against the distance of stations to the nearest fault. Red dots show the moving averages in 10 km windows.

Figure 7: Comparison between results from this study (red bars) and previous crustal anisotropy measurements from Chen et al. (2013) (black bars) and Sun et al. (2015) (green bars).

Figure 8: Spatial distribution of the splitting times of crustal anisotropy measurements. To produce the data plotted here, a smooth surface was generated by fitting the observed values, with a spatial sampling interval of 0.1° .

Figure 1

FIGURES

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