1 Seismic anisotropy beneath the incipient Okavango rift:
2 Implications for rifting initiation

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8 **Abstract**

9 This study represents the first shear-wave splitting investigation of the Oka-
10 vango rift zone (ORZ), an incipient continental rift belonging to the East
11 African rift system in northern Botswana. Analysis of broadband seismic
12 data recorded along a 750 km long profile of 22 stations traversing the ORZ
13 and adjacent Congo and Kalahari cratons and several Precambrian orogenic
14 zones reveals dominantly NE-SW fast orientations, which are parallel to both
15 the absolute plate motion direction (based on the NNR-NUVEL-1A model)
16 and the trend of most tectonic boundaries, including that of the ORZ. Spatial
17 coherence analysis of the splitting parameters and correspondence between
18 the observed fast orientations and the trend of tectonic features indicate that
19 the main source of observed anisotropy is most likely in the upper astheno-
20 sphere, probably due to simple shear associated with the relative movement
21 of the lithosphere against the asthenosphere. The presence of consistently
22 rift-parallel fast orientations and normal splitting times in the ORZ and
23 most parts of southern Africa implies that neither an upper mantle plume
24 nor small-scale convection is the dominant source for rift initiation and de-

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velopment. The first SWS measurements in the vicinity of the ORZ favor a model in which continental rifting develops in response to intra-plate relative movement of continental blocks along zones of weakness produced by ancient tectonic events.

Keywords:

Seismic anisotropy, Okavango rift zone, Shear wave splitting

1. Introduction

Continental rifting constitutes the transition from stable continental lithosphere into an ocean basin. Numerous studies suggested that continental rifting is either induced by active mantle upwelling associated with deep mantle plumes (active rifting), or by horizontal motion of plates and their interaction along plate boundaries (passive rifting; Sengor and Burke, 1978; Hill, 1991). These contrasting models of continental rupture are primarily assembled from observations of mature rift zones. In contrast, knowledge of rifting mechanisms and the associated characteristic lithospheric and asthenospheric structure and dynamics beneath incipient rift zones, such as the Okavango rift zone (ORZ) in northern Botswana, is severely limited.

The ORZ (Fig. 1) is one of the youngest continental rifts in the world with an estimated initiation of surface rupture between 120 and 40 ka (Kinabo et al., 2008) and thus is an ideal locale to study the characteristics of earliest-stage rifting mechanisms. Situated at the southwestern terminal of the East African rift system (EARS), the ORZ is superimposed upon the Neoproterozoic Damara and Rehoboth orogenic belts (McCourt et al., 2013), which are bounded toward the northwest and southeast by the Congo and

49 Kalahari cratons, respectively (Begg et al., 2009). The latter is a composite
50 terrane composed largely of the Archean Kaapvaal and Zimbabwe cratons
51 and the Archean Limpopo belt. It is encircled by the Paleo- to Mesopro-
52 terozoic Namaqua-Natal belt to the south and the Paleoproterozoic Magondi
53 belt and Rehoboth province toward the west (McCourt et al., 2013) (Fig. 1).

54 While the crustal and mantle structure and dynamics of the Kalahari
55 craton and the mature sections of the EARS have been extensively stud-
56 ied through numerous projects over the past 40 years, the ORZ has not
57 been investigated using robust structural seismological techniques. One such
58 technique is shear-wave splitting (SWS) analysis. It has long been recog-
59 nized that near-vertically incident shear waves propagating in a transversely
60 isotropic medium split into two waves with orthogonal polarization orien-
61 tations (Crampin, 1981). SWS analysis is commonly employed to quantify
62 in-situ lithospheric and asthenospheric seismic anisotropy (e.g., Silver and
63 Chan, 1991). It uses P-to-S converted phases at the core-mantle boundary,
64 including PKS, SKKS, and SKS (which are collectively called XKS), to ob-
65 tain the two splitting parameters, including the polarization orientation of
66 the fast component (fast orientation or ϕ), and the splitting time between
67 the fast and slow components (splitting time or δt). The former is an indi-
68 cator of the orientation of the anisotropic structure, and the latter measures
69 the strength of anisotropy integrated over the whole raypath from the core-
70 mantle boundary to the recording station.

71 Seismological and petrophysical studies (e.g., Silver, 1996; Debayle and
72 Ricard, 2013) have indicated that the dominant contribution to splitting pa-
73 rameters lies primarily within the upper mantle. Several mechanisms have

74 been postulated to produce observed tectonic-scale observable anisotropy.
 75 Based upon the strong correlation between splitting parameters and surfi-
 76 cial geologic features, seismic anisotropy has been proposed to exist in the
 77 lithosphere due to fabrics generated by vertically-coherent lithospheric de-
 78 formation (Silver, 1996; Silver et al., 2001), or lithosphere-scale magmatic
 79 dikes (Gao et al., 1997). Alternatively, seismic anisotropy may be induced
 80 by present-day upper-mantle flow either as a consequence of simple shear be-
 81 tween the rigid lithosphere and mobile asthenosphere, which in turn would
 82 produce anisotropy parallel to the absolute plate motion (APM), if the di-
 83 rection of the shear is the same as the APM direction (e.g., Vinnik et al.,
 84 1996; Barruol and Ismail, 2001; Gao et al., 2010; Lemnifi et al., 2015), or
 85 by small-scale mantle convection (e.g., Gao et al., 1994; Koch et al., 2012).
 86 A recent global-scale surface wave azimuthal anisotropy study (Debayle and
 87 Ricard, 2013) suggested a weak correlation between APM and fast directions
 88 for plates moving at a rate slower than 2.5 cm/yr; such a correlation increases
 89 rapidly for plates moving between 2.5 and 5.5 cm/yr, and saturates for those
 90 moving faster than 5.5 cm/yr (see Fig. 6 of Debayle and Ricard, 2013). Note
 91 that according to the no-net rotation (NNR) NUVEL-1A model (DeMets et
 92 al., 1994), the African continent moves at a rate of 2.8-3.2 cm/yr toward
 93 the northeast, and thus the APM is expected to be marginally correlated
 94 with the fast orientations if the relationship revealed by Debayle and Ricard
 95 (2013) applies precisely for Africa.

96 SWS studies have been widely conducted in numerous regions of continen-
 97 tal rifting affinities. Consistently rift-parallel fast orientations and splitting
 98 times substantially larger than the global average of 1.0 s (Silver, 1996) were

99 obtained for the Afar and Main Ethiopian (Gashawbeza et al., 2004; Kendall
100 et al., 2006; Gao et al., 2010), eastern African (Bagley and Nyblade, 2013),
101 Rio Grande (Sandvol et al., 1992; Liu et al., 2014; Refayee et al., 2014),
102 and Baikal (Gao et al., 1997) rifts. Such rift-parallel anisotropy has been
103 attributed to a number of mechanisms, including rift-parallel mantle flow
104 (Sandvol et al., 1992; Gao et al., 2010; Bagley and Nyblade, 2013), litho-
105 spheric fabrics formed by past tectonic events (Gashawbeza et al., 2004), and
106 rift-parallel lithospheric dikes (Gao et al., 1997). Rift-orthogonal anisotropy
107 observed in the flanking area of the Baikal rift was hypothesized as reflecting
108 the horizontal branches of a small-scale mantle convection system (Gao et
109 al., 1994; 1997).

110 In general, most modern continental rift zones are associated with rift-
111 parallel fast orientations. Major exceptions to this observation include the
112 Baikal rift zone and the Red Sea which has partially transitioned into an
113 oceanic spreading center. Recent SWS studies (Elsheikh et al., 2014; Lemnifi
114 et al., 2015) in northern Africa and Arabia revealed N-S fast orientations on
115 both sides of the Red Sea, which has a nearly NW-SE strike. Such N-S
116 orientated anisotropy observed in northern Africa and Arabia was proposed
117 to be the result of simple shear developed in the boundary layer between the
118 lithosphere and the asthenosphere formed by long-term northward motion of
119 the African plate relative to the asthenosphere (Elsheikh et al., 2014; Lemnifi
120 et al., 2015).

121 Splitting measurements are absent in the vicinity of the ORZ due to the
122 heretofore paucity of broadband seismic data. Previous measurements in
123 southern Africa were concentrated on the Kaapvaal and Zimbabwe cratons

124 and the Limpopo belt (Fig. 1). An investigation by Vinnik et al. (1996)
 125 at 7 stations on the Kaapvaal craton revealed NE-SW oriented fast orien-
 126 tations, which are parallel to the APM direction of Africa since Jurassic
 127 computed based on the NNR-NUVEL-1A model (DeMets et al., 1994), and
 128 were therefore interpreted to be the result of APM induced asthenospheric
 129 flow. Subsequently, Silver et al. (2001) used a much larger data set recorded
 130 at 79 sites in southern Africa and proposed that the observed anisotropy
 131 is the result of vertically-coherent lithospheric deformation. Surface-wave
 132 studies beneath southern Africa suggested that azimuthal anisotropy varies
 133 with depth, implying the presence of anisotropy both in the lithosphere and
 134 asthenosphere (Adam and Lebedev, 2012).

135 This study is part of a multi-disciplinary investigation of the incipient
 136 segments of the EARS (Gao et al., 2013). Our main objective is to present
 137 results from the first examination of mantle anisotropy using SWS in the
 138 vicinity of the ORZ in order to provide constraints on models of rift initiation
 139 and anisotropy formation. In doing so, we explore possible relationships
 140 between the observed anisotropy with predicted directions of rifting-related
 141 mantle flow, the direction of present-day plate motion, and the dominant
 142 trend of Precambrian tectonic boundaries.

143 **2. Data and Methods**

144 *2.1. Data*

145 The broadband seismic data used in the study were recorded by 17 Seismic
 146 Arrays for African Rift Initiation (SAFARI) stations that we installed and
 147 operated for a two-year period starting from the summer of 2012 (Gao et

148 al., 2013). Data recorded by five stations belonging to the 1997-99 Southern
 149 African Seismic Experiment (SASE; Silver et al., 2001) situated immediately
 150 within our study area are also used. One of the SAFARI stations, B1665,
 151 is located about 1 km away from SASE station SA65 (Fig. 2). The total
 152 length of the profile is about 750 km (Figs. 1 and 2). XKS waveforms were
 153 requested from the Incorporated Research Institutions for Seismology (IRIS)
 154 Data Management Center (DMC). The epicentral distance range for PKS,
 155 SKKS, and SKS is $120\text{-}180^\circ$, $95\text{-}180^\circ$ and $84\text{-}180^\circ$, respectively (Liu and Gao,
 156 2013). We apply a cutoff magnitude of 5.6 for events shallower than 100 km
 157 while allow for magnitudes from deeper events as low as 5.5 for the purpose
 158 of taking advantage of sharper waveforms.

159 To identify and correct for possible sensor misorientation, we apply the
 160 technique developed by Niu and Li (2011) to minimize the P-wave energy
 161 on the transverse component. We begin by manually picking the first arrival
 162 of the direct P-phase on the vertical component for events recorded by no
 163 fewer than 5 stations. The optimal orientation of the N-S component cor-
 164 responding to the minimum P-wave energy on the transverse component is
 165 found using a grid-search algorithm, and is subsequently used to correct for
 166 station orientation prior to SWS analysis.

167 *2.2. Measuring shear wave splitting parameters*

168 SWS parameters were calculated using the procedure of Liu and Gao
 169 (2013) which was based on the method of minimization of transverse energy
 170 (Silver and Chan, 1991). The seismograms were first windowed to include
 171 the theoretical arrival of the individual XKS phases and then enhanced by
 172 applying a band-pass filter with corner frequencies of 0.04-0.5 Hz. All the

173 SWS measurements were visually verified and, if necessary, various adjust-
174 ments were applied to the beginning and end times of the XKS window,
175 quality ranking, and/or band-pass filtering frequencies to ensure reliability
176 (Liu and Gao, 2013).

177 Following the evaluation criteria of Liu and Gao (2013), we ranked well-
178 defined splitting measurements as either A (outstanding) or B (good) for
179 use in the following SWS analyses. We also analyzed the data set for null
180 measurements, which are characterized by a lack of observable energy on the
181 transverse component concurrent with a strong XKS arrival on the radial
182 component. Situations wherein the backazimuth (BAZ) is parallel or per-
183 pendicular to the fast orientation, or the sub-station crust and mantle are
184 isotropic, can all result in null measurements (Silver and Chan, 1991; Liu and
185 Gao, 2013). As shown in Fig. 2, well-defined splitting parameters were ob-
186 served at all stations in the study area, suggesting the existence of pervasive
187 anisotropy. Consequently, null measurements will not be discussed below.

188 *2.3. Anisotropy depth determination*

189 The epicentral distance of the SKS, SKKS, and PKS arrivals used in the
190 study is equal to or greater than 84° , 95° , and 120° , respectively. For SKS,
191 this corresponds to a ray parameter of 6.532 s/degree or smaller, and an
192 angle of incidence of smaller than 16° at 200 km depth measured from the
193 vertical. Similarly, the maximum angle of incidence at 200 km depth for
194 the SKKS and PKS phases is about 18° and 5° , respectively. Therefore, the
195 XKS phases arrive at a steep angle of incidence and thus have an excellent
196 lateral but poor vertical resolution, resulting in uncertainties and debates in
197 the tectonic interpretations of the observed SWS measurements.

198 A technique commonly employed to estimate anisotropy depth is the in-
199 tersecting Fresnel-zone approach (Alsina and Snieder, 1995) which is based
200 on the analysis of the size for the first Fresnel zones of close-proximity XKS
201 raypaths. However, such a technique is limited to the case wherein signifi-
202 cant lateral variations in seismic anisotropy exist between nearby stations, or
203 when a station samples two regions with different splitting parameters (see
204 Lemnifi et al., 2015 for an example in Libya).

205 In this study, we apply a more generalized and quantitative extension
206 of the intersecting Fresnel zone approach based on the spatial coherency
207 of splitting parameters (Liu and Gao, 2011). This technique employs the
208 idea that the observed splitting parameters will reach the highest spatial
209 coherency if the assumed anisotropy depth is correct.

210 The optimal anisotropy depth is searched in the range of 0 to 350 km at
211 an interval of 5 km. It corresponds to the minimum spatial variation factor
212 (F_v), which is defined in Eqs. 4-7 in Liu and Gao (2011). The block size
213 (dx) used to calculate the F_v should be given a proper value due to the fact
214 that smaller dx will bring unstable F_v depth variations and uncertainties,
215 while larger dx leads to a broadened F_v curve and reduces the peak-to-peak
216 amplitude (Liu and Gao, 2011). In this study, experimental values of dx
217 from 0.16° to 0.28° with an interval of 0.02° have been employed to generate
218 the F_v curves.

219 3. Results

220 3.1. Sensor orientation correction

221 Table 1 shows the optimal orientation of the N-S component of the 3-
222 component sensor for each of the 22 stations. The magnitude of the orienta-
223 tion ranges from -16 to 5° clockwise from the north, with a simple mean of
224 $-3 \pm 1^\circ$. The overwhelmingly negative values of the resulting orientation were
225 most likely the consequences of inadequate correction of the declination of
226 the geomagnetic field, which is about -9° in the study area.

227 After the application of station orientation correction, the quality of the
228 resulting SWS measurements at stations with misoriented sensors is notice-
229 ably improved. Fig. 3 shows an example of the improvement for station
230 B07DX. When the uncorrected data are used for SWS analysis, remaining
231 XKS energy on the corrected transverse seismogram is observable, and the
232 match between the fast and slow components is poor and is accompanied
233 by weak linearity of the corrected particle motion (Fig. 3a). All of these
234 features are indicators of sensor misorientation, as discussed in Liu et al.
235 (2008). The post orientation correction measurement is substantially im-
236 proved in terms of the removal of transverse energy, the matching of the fast
237 and slow waveforms, and the linearity of the particle motions (Fig. 3b).

238 3.2. Spatial distribution of XKS splitting parameters

239 A total of 223 pairs of Quality A or B splitting parameters were obtained
240 at the 22 stations, including 73 PKS, 58 SKKS, and 92 SKS measurements
241 from a total of 91 teleseismic events (Fig. 2c). Examples for each of the three
242 XKS phases are demonstrated in Fig. 4. No systematic azimuthal variations

243 of the splitting parameters are observed (Fig. 5), suggesting that a single
 244 layer of anisotropy with a horizontal axis of symmetry is sufficient to explain
 245 the observed SWS parameters (Silver and Savage, 1994). The fast orienta-
 246 tions are dominantly NE-SW with an average value of $37.5 \pm 20.6^\circ$ (Figs. 1
 247 and 2, Table 2), which is consistent with the APM direction of the African
 248 plate in the study area based on the NNR-NUVEL-1A model (DeMets et al.,
 249 1994). The mean splitting time is 1.09 ± 0.34 s and is comparable with the
 250 global average of 1.0 s for continents (Silver, 1996), but is larger than the
 251 0.62 ± 0.02 s observed at the 59 non-null SASE stations (Silver et al., 2001).
 252 Under the assumption of a 4% anisotropy (Mainprice et al., 2000), the thick-
 253 ness of the anisotropy layer that produces the observed anisotropy is about
 254 120 km on average.

255 Based on the location of the tectonic boundaries (Begg et al., 2009; Mc-
 256 Court et al., 2013) and the characteristics of the observed splitting param-
 257 eters, we divide the measurements into 4 groups (Fig. 2a). The 7 mea-
 258 surements in Area A (Figs. 5a and 5b) have the surface projection of their
 259 ray-piercing points (calculated at the depth of 250 km) on the Congo craton.
 260 The mean splitting parameters are $56.1 \pm 24.1^\circ$ for ϕ , and 1.10 ± 0.38 s for
 261 δt . The 48 Area B measurements are in the ORZ, with corresponding mean
 262 splitting parameters of $30.3 \pm 19.3^\circ$ and 1.12 ± 0.31 s (Figs. 5c and 5d). Area
 263 C includes 83 measurements in the Magondi belt and the southern half of
 264 the Rehoboth Province (Figs. 5e and 5f). The mean splitting parameters
 265 are $27.3 \pm 17.4^\circ$ and 1.13 ± 0.40 s. The most significant spatial variation of the
 266 splitting parameters in the entire study is found near the SE end of Area C
 267 (Fig. 2a). Area D belongs to the Kalahari craton (mostly the Limpopo belt).

268 It has 85 measurements (Figs. 5g and 5h) and the mean splitting parameters
269 are $50.16 \pm 15.7^\circ$ and 1.04 ± 0.30 s.

270 3.3. Relationship with SWS studies from adjacent areas

271 All of the previous studies in adjacent areas reported their results in the
272 form of station-averaged parameters, which are largely consistent with those
273 obtained at 22 stations measured by this study (Fig. 1). Barruol and Ismail
274 (2001) used African IRIS and Geoscope permanent stations to investigate up-
275 per mantle anisotropy, including station TSUM in Namibia near the southern
276 border of the Congo craton as well as station LSZ in Zambia northeast of
277 the ORZ (Fig. 1). Both LSZ and TSUM display SWS parameters consistent
278 with our findings, i.e., a nearly NE-SW fast orientation (16° and 31° , re-
279 spectively) and small splitting time (0.73 s and 0.43 s, respectively) without
280 evidence of azimuthal variations. Similarly, relatively small splitting times
281 are observed for the SAFARI stations near the Congo craton (e.g., B14MH
282 and B13NX) while relatively larger splitting times are reported for stations
283 within the ORZ (about 1.12 s on average).

284 Among the 17 SAFARI stations, three (B01KR, B02LT, and B1665) are
285 within the study area of the SASE project (Silver et al., 2001). The re-
286 sulting SWS parameters from the three SAFARI stations are remarkably
287 consistent with those from the five SASE stations along the profile (Silver et
288 al., 2001). The average fast orientation and splitting time of the five SASE
289 stations are $53.4 \pm 14.6^\circ$ and 1.07 ± 0.25 s, while those of the three SAFARI
290 stations are $48.2 \pm 15.7^\circ$ and 1.03 ± 0.35 s. The individual SWS measurements
291 demonstrating an approximate E-W fast orientation are similar with the
292 station-averaged SWS measurements obtained by Silver et al. (2001) from

293 other SASE stations within the Limpopo belt (Figs. 1 and 2).

294 4. Discussion

295 4.1. Depth of anisotropy

296 Several conditions must be satisfied prior to applying the anisotropy-
297 depth estimation technique (Liu and Gao, 2011), the first of which is the
298 compilation of a high-quality data set of individual (rather than station-
299 averaged) splitting parameters, and the other is that there must exist a sig-
300 nificant but smooth spatial variation in the splitting parameters which should
301 simultaneously signify a single layer of anisotropy (Liu and Gao, 2011). We
302 find that the observed SWS measurements in Area C possess significant spa-
303 tial variations (Fig. 2a) and satisfy the other conditions mentioned above,
304 and can be used to estimate the depth of anisotropy.

305 The depths corresponding to the minimum variation factor on the result-
306 ing F_v curves (Figure 6) calculated using the spatial coherency technique
307 suggest that the optimal depth of the center of the anisotropic layer is ap-
308 proximately between 240 and 280 km, which is similar to the lithospheric
309 thickness based on the tomography results for southern Africa (James et al.,
310 2001). Beneath the vicinity of the ORZ, magnetotelluric studies suggested
311 a lithospheric thickness of about 180 km (Miensopust et al., 2011), and sur-
312 face wave tomography revealed a similar thickness of 180-220 km (Fishwick,
313 2010). Thus it can be safely concluded that the anisotropy originates primar-
314 ily from the upper asthenosphere beneath the ORZ, which is similar to the
315 conclusion proposed by Vinnik et al. (1996) for the Kaapvaal craton. It must
316 be mentioned, however, that due to a lack of spatial variations of the splitting

317 parameters in Areas A, B, and D, the depth of the source of the observed
318 anisotropy cannot be estimated beneath these areas; however, as discussed
319 below, although lithospheric contributions cannot be completely ruled out,
320 the observations can best be explained if they have an asthenospheric origin.

321 *4.2. Constraints on rifting mechanisms*

322 The similarity between the APM direction and the trend of most of the
323 tectonic boundaries in the study area makes it difficult to distinguish a litho-
324 spheric origin from an asthenospheric one. In addition to the results of depth
325 estimation, which favors an upper asthenospheric origin, another piece of ev-
326 idence arguing against a dominantly lithospheric origin can be found along
327 the boundary between the Zimbabwe craton and the Limpopo belt (Fig. 2),
328 where all the fast orientations are almost orthogonal (rather than parallel,
329 as expected for a lithospheric origin) to this major tectonic boundary. Ad-
330 ditionally, the ORZ, which is characterized by a relatively thin lithosphere
331 compared to the adjacent cratonic terranes based upon magnetotelluric and
332 tomographic studies (James et al., 2001; Miensoopust et al., 2011), exhibits
333 slightly larger splitting times than the Kalahari craton. One would expect
334 to observe smaller splitting times in regions with thin lithosphere if the
335 anisotropy is the result of uniform strain pervading the continental litho-
336 sphere (Silver and Chan, 1991). This interpretation is also consistent with
337 the fact that the SASE stations (Silver et al., 2001), most of which overlie
338 the thicker lithosphere of the Kalahari craton, show smaller splitting times
339 with a mean value of 0.62 ± 0.02 s (and about 1/4 of the stations show no
340 splitting) relative to the SAFARI stations (1.09 ± 0.34 s).

341 In the following, we examine a number of possible geodynamic models for

342 the initiation and development of continental rifting, under the assumption
343 that the observed anisotropy primarily reflects simple shear in an approxi-
344 mately 120 km thick layer in the upper-most asthenosphere.

345 *4.2.1. Active rifting*

346 Active rifting models (e.g., Sengor and Burke, 1978) advocate the dom-
347 inant role of mantle flow, presumably associated with a mantle plume, on
348 continental breakup. For a stationary lithosphere relative to the astheno-
349 sphere, the horizontal component of the flow system is expected to have a
350 pattern that is radiating away from the center of the plume, and for a moving
351 lithosphere, a parabolic pattern is predicted (Karato et al., 2008).

352 The existence of such a plume beneath the ORZ is possible, due to the
353 fact that the ORZ is developing directly above the proposed African super-
354 plume (Ritsema et al., 1998; Hansen et al., 2012). The expected radiating or
355 parabolic pattern of the fast orientations, however, is not observed beneath
356 the ORZ or in southern Africa (Fig. 1), leading to doubts about the existence
357 of an active mantle plume beneath the area. Such a conclusion is consistent
358 with the observations of normal mantle transition zone thickness using the
359 same dataset (Yu et al., 2015), as well as results from seismic tomography
360 studies suggesting that the Superplume beneath southern Africa is limited
361 in the lower mantle (e.g., Ritsema et al., 1998).

362 *4.2.2. Edge-driven mantle convection*

363 The thin lithosphere beneath the ORZ relative to the Congo craton could
364 induce a small-scale mantle convection system due to lateral temperature
365 variations (King and Anderson, 1998). This mechanism has been used to

366 explain the formation of zones of continental extension such as the Cameroon
 367 volcanic line (Koch et al., 2012), although other mechanisms such as three-
 368 dimensional convective instability at the base of the lithosphere (Fourel et al.,
 369 2013) have been proposed to explain volcanisms along the edge of the Congo
 370 Craton. If this process contributes significantly to the observed anisotropy,
 371 the fast orientations in the vicinity of the ORZ should be mostly orthogonal
 372 to the trend of the rift. We observed dominantly rift-parallel fast orientations
 373 and thus conclude that edge-driven small-scale convection is not the source
 374 of the observed anisotropy.

375 *4.2.3. Rifting induced by intra-plate relative motion*

376 The ORZ is developing primarily within the Damara orogenic belt be-
 377 tween the Archean Congo and Kalahari cratons. The consistently NE-SW
 378 fast orientations are largely in agreement with those observed in surrounding
 379 areas. Though the African plate has a general northeastward motion based
 380 on the NUVEL-1A model (DeMets et al., 1994), it has been suggested that
 381 South Africa is currently rotating clockwise relative to the Nubian plate,
 382 probably along the Damara belt (Malservisi et al., 2013).

383 Numerical modeling shows that topography of the lithosphere-asthenosphere
 384 boundary exerts a strong control on the direction and magnitude of stress
 385 transfer from the asthenosphere to the lithosphere (O'Neill et al., 2010). In
 386 addition, increased plate-mantle coupling beneath thick continental litho-
 387 sphere may increase plate-driving forces, surface deformation, and mantle-
 388 derived lithospheric stresses in these regions (Conrad and Lithgow-Bertelloni,
 389 2006). Therefore, contrasts in lithospheric thickness between the Congo cra-
 390 ton, the ancient orogenic belts, and the Kalahari craton can possibly lead

391 to differential basal drag forces beneath different areas, leading to spatially-
 392 varying plate motion velocities. The relative movements between the Archean
 393 cratons could rupture ancient zones of lithospheric weakness such as the
 394 Damara belt and exert a trans-tensional force upon the lithosphere, resulting
 395 in the initiation of continental rifting. This is consistent with the observa-
 396 tion that most rifts develop within weak orogenic zones (Keller et al., 1991;
 397 Vauchez et al., 1997). The spatially consistent NE-SW fast orientations ob-
 398 served in the study area as well as the rest of southern Africa are mostly
 399 parallel to the APM predicted by the NUVEL-1A model, which suggests a
 400 plate motion rate of about 3 cm/yr. This is slightly above the threshold
 401 value of 2.5 cm/yr (Debaille and Ricard, 2013), above which a moving plate
 402 is expected to exert simple shear to the asthenosphere and induce azimuthal
 403 anisotropy.

404 This model suggests a passive role of mantle flow in the initiation and
 405 development of the ORZ. Given the consistency between the APM and the
 406 fast orientations revealed by this and previous studies (Fig. 1), we propose
 407 that simple shear in the direction of the APM developed in the upper as-
 408 thenosphere contributes the bulk of the observed anisotropy. Deviations of
 409 the fast orientation observed in Area C can be explained as the modulation
 410 of flow by the topography of the bottom of the lithosphere, as proposed for
 411 North America (Fouch et al., 2000; Refayee et al., 2014) and northern Africa
 412 (Lemnifi et al., 2015). The model is consistent with surface wave studies
 413 which demonstrate the existence of asthenospheric plate-parallel polariza-
 414 tion of azimuthal anisotropy beneath southern Africa (Adam and Lebedev,
 415 2012), and with findings from a joint inversion of P-wave receiver functions

416 and SWS waveforms (Vinnik et al., 2012).

417 5. Conclusions

418 In this study, we employed recently recorded broadband seismic data in
419 the vicinity of the ORZ to conduct the first SWS investigation of the incipient
420 rift. The resulting mantle anisotropy shows a dominantly NE-SW orienta-
421 tion which is parallel to the African APM. Although significant contributions
422 from the lithosphere to the observed anisotropy cannot be completely ruled
423 out, spatial coherency analysis of the splitting parameters and correspon-
424 dence with geological features suggest that the center of the approximately
425 120 km thick anisotropic layer is located between the depth of 240 and 280
426 km, implying an asthenospheric origin of anisotropy, probably as the result
427 of simple shear in the boundary layer between the lithosphere and the as-
428 thenosphere. Comparison of the resulting SWS measurements with predicted
429 mantle flow directions originating from an active mantle plume or edge-driven
430 small-scale mantle convection suggests that neither is operating in the up-
431 per mantle beneath the incipient rift. The measurements support the rifting
432 initiation model in which differential basal drag applied to cratonic blocks
433 results in relative intra-plate movements, and leads to rifting along ancient
434 orogenic zones, which are areas of mechanical weakness.

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444 References

- 445 Adam, J.M.C., Lebedev, S., 2012. Azimuthal anisotropy beneath southern
446 Africa from very broad-band surface-wave dispersion measurements, *Geophys. J. Int.*, 191, 155-174, doi:10.1111/j.1365-246X.2012.05583.x.
- 448 Alsina, D., Snieder, R., 1995. Small-scale sublithospheric continental mantle
449 deformation: Constraints from SKS splitting observations, *Geophys. J. Int.*, 123, 431-448, doi: 10.1111/j.1365-246X.1995.tb06864.x.
- 451 Bagley, B., Nyblade, A.A., 2013. Seismic anisotropy in eastern Africa, mantle
452 flow, and the African superplume, *Geophys. Res. Lett.*, 40, 1500-1505,
453 doi:10.1002/grl.50315.
- 454 Barruol, G., Ismail, W.B., 2001. Upper mantle anisotropy beneath the
455 African IRIS and Geoscope stations, *Geophys. J. Int.*, 146, 549-561,
456 doi:10.1046/j.0956-540x.2001.01481.x.
- 457 Begg, G.C., Griffin, W.L., Natapov, L.M., O'Reilly, S.Y., Grand, S.P.,
458 O'Neill, C.J., Hronsky, J.M.A., Poudjom Djomani, Y., Swain, C.J., Deen,
459 T., Bowden, P., 2009. The lithospheric architecture of Africa: Seismic to-
460 mography, mantle petrology, and tectonic evolution, *Geosphere*, 5, 23-50,
461 doi:10.1130/GES00179.1.

- 462 Conrad, C.P., Lithgow-Bertelloni, C., 2006. Influence of continental roots and
463 asthenosphere on plate-mantle coupling, *Geophys. Res. Lett.*, 33, L05312,
464 doi:10.1029/2005GL025621.
- 465 Crampin, S., 1981. A review of wave motion in anisotropic and cracked
466 elastic-media, *Wave Motion*, 3, 343-391, doi:10.1016/0165-2125(81)90026-
467 3.
- 468 Debayle, E., Ricard, Y., 2013. Seismic observations of large-scale deformation
469 at the bottom of fast-moving plates, *Earth Planet. Sci. Lett.*, 376, 165-177,
470 doi: 10.1016/j.epsl.2013.06.025.
- 471 DeMets, C., Gordon, R.G., Argus, D.F., Stein, S., 1994. Effect of recent revi-
472 sions to the geomagnetic reversal time scale on estimates of current plate
473 motions, *Geophys. Res. Lett.*, 21, 2191-2194, doi: 10.1029/94GL02118.
- 474 Els Sheikh, A.A., Gao, S.S., Liu, K.H., Mohamed, A.A., Yu, Y., Fat-Helbary,
475 R.E., 2014. Seismic anisotropy and subduction-induced mantle fabrics be-
476 neath the Arabian and Nubian Plates adjacent to the red Sea, *Geophys.*
477 *Res. Lett.*, 41, 2376-2381.
- 478 Fishwick, S., 2010. Surface wave tomography: Imaging of the lithosphere-
479 asthenosphere boundary beneath central and southern Africa? *Lithos*,
480 120, 63-73. doi:10.1016/j.lithos.2010.05.011.
- 481 Fouch, M.J., Fischer, K., Parmentier, E.M., Wysession, M.E., Clarke, T.J.,
482 2000. Shear-wave splitting, continental keels, and patterns of mantle flow,
483 *J. Geophys. Res.*, 105, 6255-6275, doi:10.1029/1999JB900372.

484 Fourel, L., Milelli, L., Jaupart, C., Limare, A., 2013. Generation of continen-
485 tal rifts, basins, and swells by lithosphere instabilities, *J. Geophys. Res.*,
486 118, 3080-3100, doi:10.1002/jgrb.50218.

487 Gao, S., Davis, P.M., Liu, H., Slack, P.D., Zorin, Y.A., Mordvinova, V.V.,
488 Kozhevnikov, V.M., Meyer, R.P., 1994. Seismic anisotropy and mantle flow
489 beneath the Baikal rift zone, *Nature*, 371, 149-151, doi:10.1038/371149a0.

490 Gao, S., Davis, P.M., Liu, H., Slack, P.D., Rigor, A.W., Zorin, Y.A.,
491 Mordvinova, V.V., Kozhevnikov, V.M., Logatchev, N.A., 1997. SKS split-
492 ting beneath continental rift zones, *J. Geophys. Res.*, 102, 22,781-22,797,
493 doi:10.1029/97JB01858.

494 Gao, S.S., Liu, K.H., Abdelsalam, M.G., 2010. Seismic anisotropy beneath
495 the Afar Depression and adjacent areas: Implications for mantle flow, *J.*
496 *Geophys. Res.*, 115, B12330, doi:10.1029/2009JB007141.

497 Gao, S.S., Liu, K.H., Reed, C.A., Yu, Y., Massinque, B., Mdala, H., Moidaki,
498 M., Mutamina, D., Atekwana, E.A., Ingate, S., Reusch, A.M., 2013. Seis-
499 mic arrays to study African rift initiation, *Eos Trans. AGU*, 94, 213, doi:
500 10.1002/2013EO240002.

501 Gashawbeza, E.M., Klemperer, S.L., Nyblade, A.A., Walker, K.T., Kera-
502 nen, K.M., 2004. Shear-wave splitting in Ethiopia: Precambrian mantle
503 anisotropy locally modified by Neogene rifting, *Geophys. Res. Lett.*, 31,
504 L18602, doi:10.1029/2004GL020471.

505 Hansen, S.E., Nyblade, A.A., Benoit, M.H., 2012. Mantle structure beneath
506 Africa and Arabia from adaptively parameterized P-wave tomography: Im-

507 plications for the origin of Cenozoic Afro-Arabian tectonism, *Earth Planet.*
 508 *Sc. Lett.*, 319, 23-34, doi:10.1016/j.epsl.2011.12.023.

509 Hill, R.I., 1991. Starting plumes and continental break-up, *Earth Planet. Sci.*
 510 *Lett.*, 104, 398-416, doi:10.1016/0012-821X(91)90218-7.

511 James, D.E., Fouch, M.J., VanDecar, J.C., Van Der Lee, S., 2001. Tecto-
 512 spheric structure beneath southern Africa, *Geophys. Res. Lett.*, 28, 2485-
 513 2488, doi:10.1029/2000GL012578.

514 Karato, S.I., Jung, H., Katayama, I., Skemer, P., 2008. Geodynamic
 515 significance of seismic anisotropy of the upper mantle: new insights
 516 from laboratory studies, *Annu. Rev. Earth Planet. Sci.*, 36, 59-95,
 517 doi:10.1146/annurev.earth.36.031207.124120.

518 Keller, G.R., Khan, M.A., Morgan, P., Wendlandt, R.F., Baldrige, W.S.,
 519 Olsen, K.H., Prodehl, C., Braile, L.W., 1991. A comparative study of
 520 the Rio Grande and Kenya rifts, *Tectonophysics*, 197, 355-371, doi:
 521 10.1016/0040-1951(91)90050-3.

522 Kendall, J.M., Pilidou, S., Keir, D., Bastow, I.D., Stuart, G.W., Ayele, A.,
 523 2006. Mantle upwellings, melt migration, and the rifting of Africa: insights
 524 from seismic anisotropy. In: Yirgu, G., Ebinger, C.J., Maguire, P.K.H.
 525 (Eds.), *The Afar volcanic province within the East African rift system*,
 526 SP259. *Geol. Soc. Spec. Publ.*, pp. 55-72, doi:10.1016/j.epsl.2007.09.038.

527 Kinabo, B.D., Hogan, J.P., Atekwana, E.A., Abdelsalam, M.G., Mod-
 528 isi, M.P., 2008. Fault growth and propagation during incipient con-
 529 tinental rifting: Insights from a combined aeromagnetic and Shut-

530 tle Radar Topography Mission digital elevation model investigation of
531 the Okavango Rift Zone, northwest Botswana, *Tectonics*, 27, TC3013,
532 doi:10.1029/2007TC002154.

533 King, S.D., Anderson, D.L., 1998. Edge-driven convection, *Earth Planet. Sc.*
534 *Lett.*, 160, 289-296, doi:10.1016/S0012-821X(98)00089-2.

535 Koch, F.W., Wiens, D.A., Nyblade, A.A., Shore, P.J., Tibi, R., Ateba,
536 B., Tabod, C.T., Nnange, J.M., 2012. Upper-mantle anisotropy beneath
537 the Cameroon Volcanic Line and Congo craton from shear wave split-
538 ting measurements, *Geophys. J. Int.*, 190, 75-86, doi:10.1111/j.1365-
539 246X.2012.05497.x.

540 Lemnifi, A.A., Liu, K.H., Gao, S.S., Reed, C.A., Elsheikh, A.A., Yu, Y.,
541 Elmelade, A.A., 2015. Azimuthal anisotropy beneath north central Africa
542 from shear wave splitting analyses, *Geochem. Geophys. Geosyst.*, 16,
543 doi:10.1002/2014GC005706.

544 Liu, K.H., Gao, S.S., 2011. Estimation of the depth of anisotropy using spatial
545 coherency of shear-wave splitting parameters, *Bull. Seismol. Soc. Am.*, 101,
546 2153-2161, doi:10.1785/0120100258.

547 Liu, K.H., Gao, S.S., 2013. Making reliable shear-wave splitting measure-
548 ments, *Bull. Seismol. Soc. Am.*, 103, 2680-2693, doi: 10.1785/0120120355.

549 Liu, K.H., Gao, S.S., Gao, Y., Wu, J., 2008. Shear wave splitting and mantle
550 flow associated with the deflected slab beneath northeast Asia, *J. Geophys.*
551 *Res.*, 113, B01305, doi: 10.1029/2007JB005178.

552 Liu, K.H., Elsheikh, A., Lemnifi, A., Purevsuren, U., Ray, M., Refayee, H.,
 553 Yang, B., Yu, Y., Gao, S.S., 2014. A uniform database of teleseismic shear
 554 wave splitting measurements for the western and central United States,
 555 *Geochem. Geophys. Geosys.*, 15, 2075-2085, doi:10.1002/2014GC005267.

556 Mainprice, D., Barruol, G., Ismail, W.B., 2000. The seismic anisotropy of the
 557 Earth's mantle: from single crystal to polycrystal, in *Earth's Deep Interior:
 558 Mineral Physics and Tomography From the Atomic to the Global Scale*
 559 (eds S.-I. Karato, A. Forte, R. Liebermann, G. Masters and L. Stixrude),
 560 AGU, Washington, D.C., doi:10.1029/GM117p0237.

561 Malservisi, R., Hugentobler, U., Wonnacott, R., Hackl, M., 2013. How rigid
 562 is a rigid plate? Geodetic constraint from the TrigNet CGPS network,
 563 South Africa, *Geophys. J. Int.*, 192, 918-928, doi:10.1093/gji/ggs081.

564 McCourt, S., Armstrong, R.A., Jelsma, H., Mapeo, R.B.M., 2013. New U-
 565 Pb SHRIMP ages from the Lubango region, SW Angola: insights into the
 566 Palaeoproterozoic evolution of the Angolan Shield, southern Congo craton,
 567 Africa, *J. Geol. Soc. London*, 170, 353-363, doi:10.1144/jgs2012-059.

568 Miensopust, M.P., Jones, A.G., Muller, M.R., Garcia, X., Evans, R.L., 2011.
 569 Lithospheric structures and Precambrian terrane boundaries in north-
 570 eastern Botswana revealed through magnetotelluric profiling as part of
 571 the southern African magnetotelluric experiment, *J. Geophys. Res.*, 116,
 572 B02401, doi:10.1029/2010JB007740.

573 Niu, F., Li, J., 2011. Component azimuths of the CEArray stations

574 estimated from P-wave particle motion, *Earthquake Sci.*, 24, 3-13,
 575 doi:10.1007/s11589-011-0764-8.

576 O'Neill, C.J., Kobussen, A., Lenardic, A., 2010. The mechanics of
 577 continental lithosphere-asthenosphere coupling, *Lithos*, 120, 55-62,
 578 doi:10.1016/j.lithos.2010.07.008.

579 Refayee, H.A., Yang, B.B., Liu, K.H., Gao, S.S., 2014. Mantle
 580 flow and lithosphere-asthenosphere coupling beneath the southwest-
 581 ern edge of the North American craton: Constraints from shear-
 582 wave splitting measurements, *Earth Planet. Sci. Lett.*, 402, 209-220,
 583 doi:10.1016/j.epsl.2013.01.031.

584 Ritsema, J., Ni, Sidao, Helmberger, D.V., Crotwell, H.P., 1998. Evi-
 585 dence for strong shear velocity reductions and velocity gradients in
 586 the lower mantle beneath Africa, *Geophys. Res. Lett.*, 25, 4245-4248,
 587 doi:10.1029/1998GL900127.

588 Sandvol, E., Ni, J., Ozalaybey, S., Schlue, J., 1992. Shear-wave split-
 589 ting in the Rio Grande Rift, *Geophys. Res. Lett.*, 19, 2337-2340,
 590 doi:10.1029/92GL02715.

591 Sengor, A.M., Burke, K., 1978. Relative timing of rifting and volcanism
 592 on Earth and its tectonic implications, *Geophys. Res. Lett.*, 5, 419-421,
 593 doi:10.1029/GL005i006p00419.

594 Silver, P.G., 1996. Seismic anisotropy beneath the continents: Probing the
 595 depths of geology, *Annu. Rev. Earth Planet. Sci.*, 24, 385-432.

596 Silver, P.G., Chan, W.W., 1991. Shear wave splitting and subcontinental
597 mantle deformation, *J. Geophys. Res.*, 96, 16,429-16,454.

598 Silver, P.G., Savage, M., 1994. The interpretation of shear-wave splitting
599 parameters in the presence of two anisotropic layers, *Geophys. J. Int.*, 119,
600 949-963, doi:10.1111/j.1365-246X.1994.tb04027.x.

601 Silver, P.G., Gao, S.S., Liu, K.H., 2001. Mantle deformation be-
602 neath southern Africa, *Geophys. Res. Lett.*, 28, 2493-2496,
603 doi:10.1029/2000GL012696.

604 Vauchez, A., Barruol, G., Tommasi, A., 1997. Why do continents
605 break-up parallel to ancient orogenic belts?, *Terra Nova*, 9, 62-66,
606 doi:10.1111/j.1365-3121.1997.tb00003.x.

607 Vinnik, L.P., Green, R.W.E., Nicolaysen, L.O., 1996. Seismic constraints on
608 dynamics of the mantle of the Kaapvaal craton, *Phys. Earth Planet. Int.*,
609 95, 139-151, doi:10.1016/0031-9201(95)03123-5.

610 Vinnik, L., Kiselev, S., Weber, M., Oreshin, S., Makeyeva, L., 2012. Frozen
611 and active seismic anisotropy beneath southern Africa, *Geophys. Res.*
612 *Lett.*, 39, L08301, doi:10.1029/2012GL051326.

613 Yu, Y., Liu, K.H., Moidaki, M., Reed, C.A., Gao, S.S., 2015. No ther-
614 mal anomalies in the mantle transition zone beneath an incipient con-
615 tinental rift: Evidence from the first receiver function study across
616 the Okavango Rift Zone, Botswana, *Geophys. J. Int.*, 202, 1407-1418,
617 doi:10.1093/gji/ggv229.

Figure 1: A topographic map of southern Africa showing the major tectonic provinces and station-averaged shear wave splitting measurements from this (blue dots and red bars) and previous (black triangles and bars) studies (Vinnik et al., 1996; Barruol and Ismail, 2001; Silver et al., 2001). Black circles represent null measurements. The Okavango rift zone is highlighted by the grey infilled area. Green arrows represent the APM direction based on the NUVEL-1A model (DeMets et al., 1994). Major tectonic boundaries are plotted as white lines modified from McCourt et al. (2013). LB: Limpopo belt. Note that the Kaapvaal and Zimbabwe cratons and the Limpopo belt are components of the Kalahari craton. The inset shows the location of southern Africa indicated by the red rectangle and the blue lines represent the rift axes of the East African rift system.

Figure 2: Cross-section and map views of the resulting splitting parameters. (a) Fast orientations projected to the thick dashed profile in (c). Blue bars represent the station-averaged values of the fast orientation while the red bars indicate each individual measurement. RP: Rehoboth province; MB: Magondi belt. (b) Same as (a) but for splitting times. (c) Individual splitting parameters (red) plotted at the 200 km depth ray-piercing points, and station-averaged parameters (blue) plotted at the location of the stations. Grey lines delineate the tectonic provinces, and thin dashed lines within the ORZ area show active faults (Kinabo et al., 2008). SAFARI stations are shown as blue dots and SASE ones as black triangles. The inset in the upper-right corner displays the location of the study area while the one in the lower-left corner shows the distribution of earthquakes (red circles) used in the study. The green star near the SE border of the ORZ marks origin point (zero distance) in (a) and (b).

Figure 3: An SKKS measurement from station B07DX (at which the N-S component has an orientation of -11°) calculated using (A) the original and (B) station-orientation corrected dataset, respectively. (Top) Original and corrected XKS radial and transverse component seismograms. (Middle) Pre- and post-correction fast and slow waveforms and particle motion patterns. (Bottom) Contour maps of transverse component energy. The red dot on the contour map indicates the optimal splitting parameters.

Figure 4: Same as Fig. 3 but for examples of each of the XKS phases (PKS, SKKS and SKS) used in the study recorded by 3 different stations. The waveforms have been corrected for station mis-orientation.

Figure 5: Azimuthal variations of resulting fast orientations (left panels) and splitting times (right panels) for (a,b) Area A, (c,d) Area B, (e,f) Area C, and (g,h) Area D measurements. Blue squares, black triangles and red circles represent PKS, SKKS, and SKS phases, respectively. The dashed lines show the situation when the BAZ equals ϕ or $\phi + 90^\circ$, i.e., along the lines null measurements are expected (and this is why there are no well-defined measurements along the lines).

Figure 6: Anisotropy analysis for the SWS measurements in Area C based on the spatial coherence technique (Liu and Gao, 2011), resulting in an optimal depth of anisotropy between 240-280 km. The triangle on each of the curves marks the depth with the minimum variation factor.

Table 1: Resulting orientation of the N-S component

Station			AZ	No.
Name	Lat.	Lon.		
	deg.	deg.	deg.	events
B01KR	-22.238	26.718	-1	102
B02LT	-21.393	25.581	2	125
B03SL	-21.121	24.764	-6	135
B04KH	-20.474	24.514	-2	125
B05MO	-20.218	24.132	-6	131
B06OR	-19.901	23.527	-10	155
B07DX	-20.549	22.749	-11	115
B08TS	-20.164	22.459	0	84
B09NK	-19.663	22.194	-16	114
B10PP	-18.913	22.543	5	118
B11ET	-19.016	22.316	-1	112
B12SS	-18.746	22.197	3	108
B13NX	-18.579	21.994	4	100
B14MH	-18.295	21.792	-4	117
B15MW	-19.631	23.827	-9	97
B1665	-22.825	27.229	-4	130
B17CI	-19.294	22.909	-13	50
SA64	-22.969	26.202	3	34
SA65	-22.818	27.222	2	39
SA66	-21.900	26.373	-3	38
SA67	-21.886	27.274	-1	35
SA70	-21.088	26.335	-4	41

Table 2: Station-averaged splitting parameters

Station			ϕ	δt	No.
Name	Lat.	Lon.			
	deg.	deg.	deg.	s	events
B01KR	-22.238	26.718	42 ± 14	1.13 ± 0.43	15
B02LT	-21.393	25.581	52 ± 11	1.06 ± 0.24	19
B03SL	-21.121	24.764	22 ± 14	1.04 ± 0.27	21
B04KH	-20.474	24.514	20 ± 10	1.31 ± 0.45	31
B05MO	-20.218	24.132	22 ± 14	0.87 ± 0.32	16
B06OR	-19.901	23.527	37 ± 14	1.19 ± 0.26	15
B07DX	-20.549	22.749	55 ± 14	1.09 ± 0.34	8
B08TS	-20.164	22.459	57 ± 13	1.34 ± 0.32	7
B09NK	-19.663	22.194	30 ± 9	1.15 ± 0.28	3
B10PP	-18.913	22.543	24 ± 12	1.19 ± 0.36	11
B11ET	-19.016	22.316	17 ± 15	0.94 ± 0.22	4
B12SS	-18.746	22.197	78 ± 8	0.99 ± 0.40	4
B13NX	-18.579	21.994	44 ± 18	1.00 ± 0.63	3
B14MH	-18.295	21.792	15 ± 15	0.65 ± 0.20	1
B15MW	-19.631	23.827	26 ± 15	1.11 ± 0.24	11
B1665	-22.825	27.229	51 ± 22	0.73 ± 0.28	7
B17CI	-19.294	22.909	22 ± 13	0.90 ± 0.18	6
SA64	-22.969	26.202	56 ± 17	0.85 ± 0.41	3
SA65	-22.818	27.222	45 ± 15	1.02 ± 0.31	5
SA66	-21.900	26.373	61 ± 12	1.09 ± 0.15	10
SA67	-21.886	27.274	59 ± 7	0.99 ± 0.24	6
SA70	-21.088	26.335	48 ± 13	1.13 ± 0.25	17