1 **3D imaging of subducting and fragmenting Indian continental lithosphere**
2 **beneath southern and central Tibet using body-wave finite-frequency**
3 **tomography**

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Abstract

We perform a finite-frequency tomographic inversion to image 3D velocity structures beneath southern and central Tibet using teleseismic body-wave data recorded by the TIBET-31N passive seismic array as well as waveforms from previous temporary seismic arrays. High-velocity bodies dip $\sim 40^\circ$ northward beneath the Himalaya and the Lhasa Terrane. We interpret these high-velocity anomalies as subducting Indian Continental Lithosphere (ICL). The ICL appears to extend further north in central Tibet than in eastern Tibet, reaching 350 km depth at $\sim 31^\circ\text{N}$ along 85°E but at $\sim 30^\circ\text{N}$ along 91°E . Low P- and S-wave velocity anomalies extend from the lower crust to ≥ 180 km depth beneath the Tangra Yum Co Rift, Yadong-Gulu Rift, and the Cona Rift, suggesting that rifting in southern Tibet may involve the entire lithosphere. The anomaly beneath Tangra Yum Co Rift extends down to about 180 km, whereas the anomalies west of the Yadong-Gulu Rift and east of the Cona Rift extend to more than 300 km depth. The low-velocity upper mantle west of the Yadong-Gulu Rift extends furthest north and appears to connect with the extensive upper-mantle low-velocity region beneath central Tibet. Thus the northward-subducting Indian Plate is fragmented along north-south breaks that permit or induce asthenospheric upwellings indistinguishable from the upper mantle of northern Tibet.

46 **1. Introduction**

47 Four distinct models for the accommodation of convergence between the Indian
48 and Eurasian continents are: 1) underthrusting of Indian lithosphere (e.g., Argand,
49 1924; Ni and Barazangi, 1984), 2) crustal injection and mantle-lithosphere subduction
50 (e.g., Zhao and Morgan, 1987), 3) uniform viscous thickening of the Tibetan crust
51 (e.g., Dewey and Burke, 1973; England and Houseman, 1986), and 4) convective
52 removal of thickened mantle lithosphere and associated plateau uplift (e.g., Molnar et
53 al., 1993). The underthrusting model posits that the Indian lithosphere slides directly
54 beneath the Tibetan crust, without an intervening Asian asthenospheric wedge. The
55 crustal-injection model predicts a significant asthenospheric wedge between the
56 subducting Indian mantle lithosphere and the Tibetan lithosphere. The
57 viscous-thickening model assumes that the thickened Tibetan lithosphere is due to
58 accordion thickening of the Indian and Asian lithospheres south and north of the
59 Indus-Yarlung Suture (IYS). Finally, the convective-removal model attempts to
60 explain the rather uniform uplift of the entire Tibetan Plateau by attributing it to the
61 loss of a continental lithospheric root after earlier thickening.

62 Over the last two decades a large number of temporary seismic arrays have been
63 operated across the Himalayas and most of Tibet. Many geophysical observations
64 have suggested that the Indian Continental Lithosphere (ICL) underthrusts southern
65 Tibet south of the Bangong-Nujiang Suture (BNS), including observations of high Pn
66 velocity and low Sn attenuation (e.g., Ni and Barazangi, 1983; McNamara et al.,
67 1995), a doublet of converted Ps phases in the lower crust and at the Moho (e.g.,

68 Owens and Zandt, 1997; Kind et al., 2002; Schulte-Pelkum et al., 2005; Nabelek et al.,
69 2009; Zhao et al., 2010; Zhang et al., 2014), shear-wave splitting features (e.g., Huang
70 et al., 2000; Fu et al., 2008; Chen et al., 2010a; Zhao et al., 2014), fast surface-wave
71 velocity anomalies (e.g., Priestley et al., 2008; Acton et al., 2010; Chen et al., 2010b;
72 Fu et al., 2010; Jiang et al., 2011; Ceylan et al., 2012; Agius and Lebedev, 2013), fast
73 body-wave velocity anomalies (e.g., Tilmann et al., 2003; Li et al., 2008; He et al.,
74 2010; Hung et al., 2011; Liang et al., 2012; Zhang et al., 2015), and gravity anomalies
75 (Jin et al., 1996; He et al., 2014). However, due to non-uniform seismic station
76 coverage and the imprecision of existing tomographic images of the Indian and
77 Tibetan lithosphere beneath Tibet, the geometry, northern extent and lateral continuity
78 of the Indian lithosphere have been much debated.

79 Significant east-west lateral variations in the geometry and the thickness of the
80 ICL have been inferred in recent studies. Zhou and Murphy (2005) proposed a
81 wholesale underthrusting of ICL with a weakened zone between 84°-92°E. Liang and
82 Song (2006) found a broad N-S zone of low Pn velocity (~8.0 km/s) along 89°-90°E.
83 From the eastern Himalayan syntaxis to the Karakoram Fault, according to Li et al.
84 (2008), the Indian lithosphere underthrusts Tibet at an increasingly shallow angle and
85 reaches progressively further to the north (Figure 1). However, other regional
86 tomographic images show high velocities extending further north toward the Jinsha
87 Suture (JS) in the easternmost part of the plateau (Liang et al., 2012; Nunn et al.,
88 2014). Bao et al. (2015) suggested that the ICL is underthrusting horizontally under
89 the western and eastern Tibetan Plateau, but it is steeply subducting under the central

90 plateau. Furthermore, receiver-function images (e.g., Kumar et al., 2006; Zhao et al.,
91 2010) as well as body- and surface-wave tomographic models (e.g., Li et al., 2008;
92 Ren and Shen, 2008; He et al., 2010; Liang et al., 2011; Ceylan et al., 2012; Nunn et
93 al., 2014) suggest significant east-west lateral variations in the dipping geometry of
94 the ICL.

95 In a similar fashion, west-east lateral variations at the surface of the plateau are
96 well known, particularly the north-south trending rifts of southern Tibet (e.g., Molnar
97 and Tapponnier, 1978; Ni and York, 1978; Armijo et al., 1986). Minor N-S diking and
98 associated east-west extension initiated by 19 Ma (Mitsuishi et al., 2012), resulting in
99 the north-south trending extensional grabens that remain active today (Figure 1, from
100 west to east, Tangra Yum Co Rift (TYR), Punqu Xianza Rift (PXR), Yadong Gulu Rift
101 (YGR), and Cona Rift (CR)). These rifts are a manifestation of east-west extension of
102 southern Tibet, roughly perpendicular to the India-Eurasia convergence direction.
103 Whether these rifts are restricted to the upper crust or whether they involve the entire
104 lithosphere is still debated (e.g., Masek et al., 1994; Yin, 2000; Tian et al., 2015).
105 Crustal-scale mechanisms proposed to explain the rifting process include
106 lower-crustal east-west flow (e.g., Molnar et al., 1981; Royden et al., 1997), oblique
107 convergence or basal shear associated with the underthrusting ICL (e.g., McCaffrey
108 and Nabelek, 1998), and inherited architectures (Faizabad and Munger-Saharsa
109 basement ridges, Figure 1) in the underthrust Indian crust (e.g., Godin and Harris,
110 2014). Lithospheric-scale processes have also been invoked, including gravitational
111 collapse following the attainment of maximum elevation (e.g., Molnar and

112 Tapponnier, 1978; Liu and Yang, 2003), deformation facilitated by mantle-lithosphere
113 delamination (Ren and Shen, 2008), and fragmentation of the ICL (e.g., Yin, 2000;
114 Liang et al., 2012; Chen et al., 2015).

115 Liang et al. (2012) discussed the geometry of the ICL and its west-east variation,
116 but that paper mainly focused on eastern Tibet based on the data coverage provided by
117 the ASCENT project. In this paper we add data from the TIBET-31N, a 600-km-long
118 west-east passive-source linear seismic array that ranges from Cuoqin to Nam Tso and
119 was operated by the Institute of Geology and Geophysics, Chinese Academy of
120 Sciences from September 2009 to November 2010 (Chen et al., 2015). Our 48
121 temporary seismographs were equipped with Guralp CMG-3ESP sensors and
122 Reftek-72A/130 data loggers. We also recorded a 10-station profile across the
123 northern YGR (Zhang et al., 2013) (Figure 1). We include all other currently available
124 seismic data to provide a body-wave travel-time tomography model that is clearly
125 improved over earlier work (Figures S1 and S2). We use our new model to address the
126 west-east variations of upper-mantle structures in southern and central Tibet.

127

128 **2. Data**

129 Linear seismic arrays provide 2D images that can be used to understand the
130 collision process that formed the high plateau but they do not provide a complete
131 picture of the 3D structure of the crust and mantle. By combining the data from
132 several 2D arrays to provide broad lateral coverage, 3D teleseismic tomography
133 images can illuminate the upper-mantle structures of southern Tibet. In this study we

134 combine data from eight temporary seismic arrays deployed over the past twenty
135 years in central and southern Tibet: INDEPTH II (12 stations, Kosarev et al., 1999)
136 and III (42 stations, Huang et al., 2000), BHUTAN (5 stations, Velasco et al., 2007),
137 Namche Barwa (69 stations, Sol et al., 2007), Hi-CLIMB (218 stations, Nabelek et al.,
138 2009), HIMNT (28 stations, Schulte-Pelkum et al., 2005), RISE (13 stations, Jin et al.,
139 2009) and TIBET-31N (57 stations, Chen et al., 2015), as well as the Global Seismic
140 Network Lhasa station (LSA) for a combined dataset of 445 stations (Figure 1). The
141 resolution of the body-wave tomography for upper mantle structures beneath the
142 Lhasa Terrane is dramatically improved by including our TIBET-31N stations
143 (Figures S1 and S2).

144 We selected earthquakes for which direct P (or S) were recorded by at least 4
145 stations from epicentral distances of 30-90° (to minimize interference from the mantle
146 transition zone and core-mantle boundary). PcP and ScS phases with shorter
147 epicentral distances were also identified to improve the data coverage (Figure 1,
148 inset). Because our dataset spans nearly 20 years of campaign data operated at
149 different epochs, we use the permanent GSN station LSA as a common reference
150 when measuring differential travel times to make inter-network measurements
151 possible. Because different frequencies sample different spatial regions along the ray
152 paths (e.g., Dahlen et al., 2000), two different frequency bands of waveform data were
153 extracted separately for P- and S-waves in order to apply a finite-frequency
154 travel-time tomography algorithm. We measured the arrival times of P and S phases in
155 high and low-frequency bands (0.5-2 Hz and 0.1-0.5 Hz for the P, 0.1-0.5 Hz and

0.05-0.1 Hz for the S), and obtained differential travel times using the AIMBAT software package (Lou et al., 2013) based on stacking-correlation (Pavlis and Vernon, 2010) and multi-channel cross-correlation (MCCC) (VanDecar and Crosson, 1990). We used 109,002 high-frequency and 34,637 low-frequency P-wave travel times and 23,614 high-frequency and 20,956 low-frequency S-wave travel times in our inversions. Lateral P- and S-wave velocity variations are directly indicated by azimuthally averaged differential travel times at each station along the TIBET-31N profile (Figure 2, a and b). These differential travel times already show a first-order feature of our final model: increased delays are seen at stations from 88-90°E, compared to stations further west and further east.

3. Method

We used finite-frequency seismic tomography to invert for P- and S-wave crustal and upper-mantle velocity structures beneath south-central Tibet. The methodology has been thoroughly discussed and routinely used in previous works (e.g., Dahlen et al., 2000; Hung et al., 2011; Liang et al., 2011), so is not described in detail here. Since only differential travel times are used in the inversion, only lateral velocity contrasts can be resolved. We present all results as % differences in V_p and V_s from an average velocity in our model at each depth.

3.1 Model parameterization and inversion

We parameterized the volume beneath our study area (20° in longitude, 14° in latitude, and 900 km in depth) with a regular $65 \times 65 \times 33$ grid centered at 90°E,

178 29°N. This model is large enough to include most of the differential travel-time
 179 sensitivity kernels in our model space, and has a cell spacing of $\sim 0.31^\circ$ longitude,
 180 $\sim 0.22^\circ$ latitude, and 28 km depth. Every cell has uniform velocity during the
 181 inversion. With this parameterization, travel time equations can be written as:

$$182 \quad d_i = G_{il} m_l \quad (5)$$

183 where d_i is the i th differential travel time, G_{il} is the differential value of the integrated
 184 volumetric kernels of the i th event contributing to the l th node, and m_l is the model
 185 parameter at the l th node. We resolved the inversion problem using a standard damped
 186 least-squares method with smoothing:

$$187 \quad \hat{m} = (G^T G + \theta^2 I + \eta D^T D)^{-1} G^T d \quad (6)$$

188 where I is the identity matrix and D are first-order Laplacian vectors. The damping
 189 parameter θ is determined empirically by examining the data-variance reduction
 190 versus the L2 norm of inverted models represented by a trade-off curve (Figure S3).
 191 We chose the damping parameter that yields an optimum variance reduction and a
 192 relatively small model norm. Smoothing parameter η is also picked empirically to get
 193 a smoothly variable model. The model discussed in this paper used a damping and a
 194 smoothing parameter that yields a variance reduction of $\sim 69\%$ for P phases and $\sim 68\%$
 195 for S phases. The mean value and standard deviation of the input travel times are
 196 0.0038 and 0.4576 for P phases. In the case of S phases, the mean value and standard
 197 deviation of the input travel times are 0.0008 and 1.0004. After the inversion, the
 198 mean value and standard deviation of the travel times are changed to -0.0014 and
 199 0.2522 for P phases and -0.0023 and 0.5642 for S phases (Figure 3). And after the

200 inversion, the final travel-time residuals in both the high- and low-frequency bands
201 were much smaller when compared with the input differential travel times (Figures 3,
202 S4 and S5)

203 **3.2 Crustal and elevation correction**

204 The raypaths of teleseismic body waves are more vertical near the surface.
205 Therefore shallow velocity structures are poorly constrained by our relative
206 travel-time data. In teleseismic travel-time tomography crustal corrections are needed
207 in order to reduce the tradeoff between crustal and mantle velocity heterogeneities. In
208 order to correct for crustal effects, we calculated frequency-dependent crustal
209 corrections for each event-station record from the synthetic crustal response of an
210 incoming plane wave beneath that station (Yang and Shen, 2006). We tested several
211 crustal models including CRUST2.0 (Bassin et al., 2000), SEAPS (Sun et al., 2004;
212 Sun et al., 2008), and CRUST1.0 (Laske et al., 2013) (Figures S6, S7, S8 and S9), but
213 we did not find any major differences in large-scale features (>200 km) (Liang et al.,
214 2012). We used the crustal structure from SEAPS where available, because it is a
215 body-wave derived model and should therefore be more comparable with our results.
216 For the few stations where SEAPS is not defined we used CRUST2.0 (Figure S6). To
217 minimize the influence of poor crustal-structure resolution we did not include the
218 kernels of the top 50 km in the inversion. Instead we used crustal corrections and an
219 additional free term for each station, which was incorporated into the inversion to
220 absorb travel-time shifts caused by remaining shallow velocity heterogeneities (Figure
221 S10). We also tested the effect of pushing the travel-time anomalies into shallow

222 depths by using a smaller damping factor for the top three layers of the model grids in
223 the inversion. When we performed this test we could not see a significant difference
224 between the upper mantle images (Figures S11 and S12). These tests demonstrated
225 our final models should not be significantly influenced by unmodeled shallow
226 structures.

227 **3.3 Resolution tests**

228 3D resolution tests were performed to evaluate the data coverage and the ability of
229 the inversion to recover upper-mantle structures (Figures 4 and 5). Synthetic travel
230 times were computed by multiplying the G-matrix with different input velocity
231 models: $\Delta t_{syn} = G \cdot \Delta c_{syn} + t_{noise}$, where Δc_{syn} represents synthetic velocity perturbations,
232 G is the differential value of the integrated volumetric kernels, Δt_{syn} represents the
233 final synthetic travel-times, and t_{noise} represents Gaussian random noise with a
234 standard deviation of 0.1 s, which is about 10% of the typical differential travel-time
235 span for a single event. Inversions were then performed using the same damping and
236 smoothing parameters as those used in our inversion of the real data.

237 Checkerboard resolution tests were conducted using input anomalies with a
238 diameter of 6 grid cells (~170 km) in all three dimensions (Figure 4). The maximum
239 input velocity perturbation of our checkerboards is $\pm 2\%$ for the P-wave and $\pm 4\%$ for
240 the S-wave. The magnitude decreases from the center of each checkerboard cell to
241 zero at the boundary as a cosine function. Our checkerboard resolution tests show that
242 input anomalies are recovered reasonably well in regions with good ray coverage.
243 However, anomaly intensities are strongly damped during the inversion to about 1/3

244 to 1/2 of the input anomaly intensity, so we do not discuss absolute anomaly
245 amplitudes in this paper.

246 In addition to the checkerboard tests we carried out custom resolution tests to
247 demonstrate our ability to recover two distinct features of our final tomographic
248 images which become important to our later interpretations, namely the northward
249 dipping high-velocity zone and an overlying low-velocity region (Figure 5). We also
250 carried out a series of resolution tests of the high-velocity zone using different depths
251 to understand the degree of vertical smearing. For both P- and S-waves, the
252 north-dipping high-velocity zone is not a shallow anomaly and must extend to > 350
253 km depth (Figures S13 and S14).

254

255 **4. Results**

256 By combining all available travel-time data we obtain comprehensive 3D images
257 of the upper mantle beneath southern and central Tibet (Figures 6, 7 and 8) that
258 provide new insights into the geometry of the ICL. A prominent high-velocity
259 anomaly is observed along the Himalayas dipping $\sim 40^\circ$ northward from ~ 80 -170 km
260 depth. We trace this anomaly from $\sim 84^\circ\text{E}$ to 92°E , the full longitudinal range of our
261 good data coverage. This high-velocity anomaly steepens from west to east, as it
262 reaches 350-km depth at $\sim 31^\circ\text{N}$ along 85°E (Figure 7a, section A-A') but at $\sim 30^\circ\text{N}$
263 along 91°E (section E-E'). There exists a lateral gap in the high-velocity structure at
264 $\sim 89^\circ\text{E}$ to 90°E (ICL shown in Figure 7a D-D', Figure 7b F-F' and Figure 8). We
265 interpret this high-velocity body as subducting ICL following previous P-wave

266 tomography results (Li et al., 2008) (Figures 1 and S15). Based on the dip of this
267 high-velocity anomaly, we use ‘subduction’ rather than ‘underthrusting’ to describe
268 the geometry of the ICL. We also note a possible thinning of the subducting ICL in
269 the down-dip direction at ~180-220 km depth (Figure 7a, sections A-A’, B-B’).

270 A continuous low-velocity anomaly lies beneath the IYS in the uppermost mantle
271 in both the P- and S-wave images (LV or LVBI: Low-Velocity Beneath the IYS,
272 Figures 6, 7, S16 and S17). This anomaly reaches only ~100-150 km depth beneath
273 the TYR whereas it reaches up to ~300 km around the YGR and the CR (Figures 7b,
274 section G-G’). This feature has previously been interpreted as fragmentation of the
275 ICL underneath southern Tibet (Liang et al., 2011; Liang et al., 2012) or alternatively
276 as delamination of the ICL beneath eastern Tibet (Ren and Shen, 2008).

277 Beneath the northern Lhasa Terrane both our P- and S-wave images show
278 consistently high velocities in the upper mantle (HV or HVBNL: High-Velocity
279 Beneath Northern Lhasa terrane, Figures 6, 7, 8, S16 and S17) except in the region
280 close to 90°E (LVBYG: Low-Velocity Beneath the Yadong-Gulu Rift, D-D’ in Figure
281 7a, and Figure S17), where there is significant low S-wave velocity and moderately
282 low P-wave velocity. In the Qiangtang Terrane, north of the BNS, both the P- and
283 S-wave images show strong low-velocity anomalies in the upper mantle (LVBNT:
284 Low-Velocity Beneath the Northern Tibet, Figure S14) albeit with relatively low
285 resolution due to less data coverage on the northern edge of our study region. The
286 low-velocity upper mantle beneath northern Tibet is in good agreement with prior
287 body- and surface-wave tomography results (e.g., Wittlinger et al., 1996; Tilmann et

288 al., 2003; Li et al., 2008; Chen et al., 2010b; Ceylan et al., 2012; Liang et al., 2012;
289 Agius and Lebedev, 2013). As in these prior studies, we find it difficult to determine
290 the depth extent of the low-velocity region.

291 The low-velocity anomaly close to 90°E (LVBYG) is connected to similar
292 low-velocity bodies beneath the IYS and central-northern Tibet (D-D' in Figure 7a,
293 Figure S17). This anomaly might indicate detachment of the Indian lithosphere (Ren
294 and Shen, 2008) or it might represent a gap in the ICL that connects the shallower
295 low-velocity body in southern Tibet (LVBI) to the much larger low-velocity regions
296 in the upper mantle of central-northern Tibet (LVBNT). We note the correspondence
297 of the LVBNT with a region of high-frequency Sn blockage (Ni and Barazangi, 1983;
298 McNamara et al., 1995), whereas the LVBYG is aligned with the southernmost extent
299 of the region of inefficient high-frequency Sn propagation (Barron and Priestley,
300 2009).

301

302 **5. Discussions**

303 **5.1 Underthrusting, subduction, or detachment of Indian continental** 304 **lithosphere?**

305 At least three different possible geodynamic scenarios for accommodating
306 convergence of India and Asia have been proposed: (1) horizontal underthrusting of
307 the entire Indian lithosphere beneath Tibet as an integral body, (2) subduction of the
308 Indian lithosphere at intermediate angles analogous to oceanic subduction (of either
309 the entire ICL or only the Indian mantle lithosphere after delamination from all or part

310 of the Indian continental lower crust), and (3) detachment (break-off) of a portion of
311 Indian lithosphere followed by vertical sinking into the deeper mantle. All three may
312 have been active at different times during orogenic evolution and all three may now
313 characterize different locations beneath the Tibetan Plateau (e.g., DeCelles et al.,
314 2002). Older seismic data from the Tibetan Plateau were widely interpreted to show a
315 clear dichotomy in the uppermost mantle between faster wave speeds beneath
316 southern Tibet and slower wave speeds beneath northern Tibet (e.g., Ni and
317 Barazangi, 1983; McNamara et al., 1995). The seismically faster and presumably
318 colder/stronger mantle beneath southern and central Tibet was widely interpreted as
319 the ICL, underlying beneath the Tibetan crust.

320 More recent work has attempted to identify west-east variations in the northern
321 limit of the ICL. Li et al. (2008) and Zhao et al. (2010) suggest the ICL extends
322 further north to the west of 89°E, whereas Liang et al. (2012) found that the ICL
323 extends further north to the east of 92°E. Nunn et al. (2014) followed Agius and
324 Lebedev (2013) in suggesting horizontal underthrusting in central Tibet (~88°E) as far
325 north as the BNS with moderate-angle subduction occurring in eastern Tibet (~93°E).
326 West-east differences in seismic velocity and the apparent northward extent of the ICL
327 could correspond to an irregular northern boundary along which a portion of the ICL
328 may have detached at an earlier time. They may also correspond to complex processes
329 of active delamination or detachment or to velocity variations inherited from the
330 Indian craton. Finally, ICL and Asian lithosphere may be more difficult to
331 differentiate further north. Because we have included more data (Figure 1) and so

improved our resolution over previous models (Figures S1 and S2), we feel confident about the existence of the P- and S-wave high-velocity body beneath southernmost Tibet that extends down to ~400 km depth with ~40° dip angle (ICL, Figures 6, 7, 8), the high-velocity body beneath the northern Lhasa terrane (HV or HVBNL, Figures 6, 7 and S17), and the low-velocity body at 100 km depth beneath the IYS (LV or LVBI, Figures 6 and 7). Horizontal underthrusting of the coherent ICL directly beneath the present Tibetan crust seems precluded everywhere from 85°E (AA') to 91°E (F-F') by the existence of the LVBI that is seen beneath the IYS in all our south-north cross-sections. This is consistent with the inefficient propagation of high-frequency Sn, presumably due to a negative velocity gradient directly below the Moho as far south as the IYS (Barron and Priestley, 2009). The geometry and nature of the deeper high-velocity body (ICL) then determines our interpretation as to whether the ICL is better described as subducting or detaching body.

Our cross-sections A-A' and B-B' (Figure 7a) show a continuous north-dipping ICL in a geometry strongly suggestive of subduction. Our tomography lacks the resolution within the crust to distinguish between subduction of the entire ICL or delamination and subduction only of the lower ICL, so we follow recent receiver-function interpretations of north-dipping converters in the upper mantle beneath southern Tibet as implying delamination of the Indian mantle lithosphere from the Indian lower crust (Kosarev et al., 1999; Shi et al., 2015). Delamination is also supported by numerical modeling that recognizes that if Indian upper crust is underthrust beneath Tibet, the remaining Indian lower lithosphere is denser than the

underlying mantle and will delaminate (Capitanio et al., 2010). The length of our proposed north-dipping subducting ICL is at a minimum ~550 km assumed to be a rigid body (Figure 7, profile B-B'). If we assume that the ICL has maintained a constant convergence velocity with respect to Tibet (currently ~20 mm/yr) across the Himalayas based on GPS measurements (Bettinelli et al., 2006), we would estimate that the current stage of continental subduction started at least ~25-30 Ma ago.

Thus far our interpretation has ignored the HVBNL. It is possible that the HVBNL represents remnant Tibetan lithosphere associated with the late Precambrian basement of the Lhasa Block (Zhu et al., 2011), in which case its presence may have little bearing on the modern geometry or earlier evolution of the subducting ICL. However, if the HVBNL represents a separate piece of the ICL, then the ICL must have underthrust horizontally beneath Tibet nearly to the BNS at an earlier stage of continental collision to emplace the HVBNL (Figure 9a), before the modern subduction geometry was established. In this scenario, the 170 km difference between the minimum length of the entire ICL on cross-section B-B' of Figure 7a (~550 km) and the distance of the HVBNL from the MFT (~380 km) provides a minimum estimate for the initiation of delamination-subduction: ~10 Ma at a convergence rate of ~20 mm/yr. If, as widely believed (e.g., Chung et al., 2009), it was the onset of roll-back or delamination of underthrust ICL that triggered widespread post-collisional potassic volcanism across southern Tibet from ~25-10 Ma, then in the scenario of Figure 9a the subducting ICL should now continue an additional ~300 km down-dip to a depth of ~600 km. These depths are below the limit of our study.

376 However, a deeper high-velocity anomaly body has been observed in previous
377 tomographic studies (Li et al., 2008). Possible lateral necking of the subducting ICL at
378 a depth of about 180-220 km (Figure 7a, e.g. B-B', C-C') could represent ongoing
379 detachment of the ICL (Figure 9a, bottom panel).

380 In contrast to A-A' and B-B' (Figure 7a), which are suggestive of a subduction
381 geometry of the down-going ICL, the anomalies present in cross-sections C-C' and
382 E-E' appear more symmetric north and south of the LVBI. This pattern suggests the
383 presence of a detachment of thickened ICL caused by Rayleigh-Taylor gravitational
384 instability (Figure 9b). In this scenario, the high-velocity anomalies at >200 km depth
385 beneath the IYS that correlate with the thickest crust (~80 km) in Tibet would
386 represent detached ICL. When considering Rayleigh-Taylor instability, the total
387 volume of the ICL should be conserved during local thickening. Assuming subduction
388 of ~100 km-thick Indian mantle lithosphere at a constant rate of ~20 mm/yr, the input
389 area per year along the B-B' profile will be $2.0 \times 10^{-3} \text{ km}^2/\text{yr}$. The area of the gap
390 between the ICL and HVBNL is $\sim 1.6 \times 10^4 \text{ km}^2$ and the area for the detached body is
391 about $4 \times 10^4 \text{ km}^2$. The difference between these two areas is $\sim 2.4 \times 10^4 \text{ km}^2$, which
392 would have taken ~12 Ma to accumulate by underthrusting of ICL. This timing seems
393 far too late to explain the potassic volcanics erupted from ~25-10 Ma (Chung et al.,
394 2009). Unless an earlier episode of instability led to detachment of a similar sized
395 body that is now below 400 km depth, a depth invisible to our tomography (e.g. as
396 shown by DeCelles et al., 2002), it is hard to reconcile our tomographic results with a
397 convective removal process that led to formation of the observed mantle melts.

398 The P-wave tomographic image of the upper mantle beneath central Tibet from
399 INDEPTH III data (Tilman et al., 2003) revealed a sub-vertical high-velocity zone
400 from about 100 km to 400 km depth, south of the BNS, interpreted as “downwelling”
401 (subducting or detaching) ICL. Our updated 3D images show a high-velocity body at
402 a similar location (Figure 7, profile D-D’, part of our HVBNL), but with less
403 west-east lateral continuity than implicitly inferred by Tilman et al. (2003) and
404 disconnected from our proposed subducting ICL.

405 Our results show significant low-velocity anomalies in the upper mantle beneath
406 southern Tibet (Figures 6 and 7). One body is centered at 90°E and the other is
407 centered at 94°E, which confirms the geometry and depth extent of the low P- and
408 S-wave velocities found by previous regional finite-frequency tomography studies
409 (Ren and Shen, 2008; Liang et al., 2011; Liang et al., 2012). Our results suggest that
410 the current ICL is not uniformly distributed at depth and has significant
411 orogen-parallel variations in both P- and S-wave velocities. The geometry of the ICL
412 is likely more complex than suggested by simple interpretations of the 2D,
413 approximately south-north, seismic profiles in southern Tibet.

414 **5.2 Inherited lateral variations of Indian lithosphere**

415 We cannot fully rule out the possibility that the upper-mantle low-velocity
416 anomalies beneath southern Tibet are due to compositional variations within the ICL
417 that either represent preserved Precambrian terrane boundaries or were formed during
418 a Permo-Triassic rifting event (Liang et al., 2012). Surface wave tomography of the
419 Dharwar Craton, south India, shows lateral inhomogeneity from 40-100 km depth up

420 to $\pm 2\%$ over length scales of $\sim 200\text{km}$ (Borah et al., 2014). This amount of lateral
421 inhomogeneity is enough to explain a significant part of our observed variations in
422 velocity (Figure 7). Caldwell et al. (2013) discuss along-strike changes in crustal
423 thickness entering the Himalayan subduction zone of $>10\text{ km}$. Three $\sim$ south-north
424 paleotopographic subsurface ridges mapped beneath the Gangetic foreland basin, and
425 possibly beneath the Himalaya (Figure 1) (Godin and Harris, 2014), probably also
426 represent along-strike crustal differences. However, given the uncertainty as to how
427 much of the Indian crust is subducted, it is unclear whether or not these along-strike
428 changes directly contribute to our observed velocity perturbations. Given the
429 likelihood that the ICL has undergone substantial deformation during subduction, we
430 conclude that the low-velocity anomalies we observed beneath southern Tibet are
431 most likely the consequence of recent fragmentation of the Indian lithosphere (Liang
432 et al., 2012). The west-east variations of the Indian lithosphere could facilitate the
433 fragmentation process (Godin and Harris, 2014), since concentrations of strain
434 occurring at changes in lithospheric thickness or density could induce changes in
435 subduction angle (Li et al., 2008; Chen et al., 2015) and trigger the proposed
436 lithospheric tearing (Liang et al., 2012). We cannot distinguish this from other
437 possible causes of lithospheric fragmentation, including the curvature of the
438 Himalayan subduction front (McCaffrey and Nabelek, 1998), gravitational instability
439 of the thickened ICL (Houseman and Molnar, 1997), and dynamics related to the
440 subduction of the Indian plate beneath southeastern Tibet and Burma..

441 **5.3 Possible link between west-east extension and the fragmentation of the ICL**

442 Our low-velocity anomalies in the shallow mantle are spatially correlated with
443 (from west to east) the Tangra Yum Co, Yadong-Gulu, and Cona Rifts (Figure 1),
444 down to depths of 100-150 km beneath the TYR and to ~400 km beneath the YGR
445 and CR. The low-velocity anomaly near 90°E connects with the low-velocity
446 anomalies beneath the IYS and north of the BNS, agreeing well with other images of
447 alternating high and low velocities beneath southern Tibet (Liang and Song, 2006;
448 Hearn et al., 2011; Wang et al., 2013). The N-S trending uppermost mantle
449 low-velocity feature beneath the YGR is also observed in the Pn tomography images
450 (Liang and Song, 2006). The Pn tomography image of Hearn et al. (2011) showed the
451 west-east variability of Pn velocities beneath the Himalayas and southern Tibet,
452 indicating that the subducted ICL is not a homogenous body. These north-south
453 trending low-velocity anomalies have previously been interpreted as evidence of
454 fragmentation of the ICL (Liang et al., 2011; Liang et al., 2012), and may also act as
455 channels for asthenospheric flow between the shallower low-velocity zone in southern
456 Tibet and the much larger low-velocity zone in central-northern Tibet (Liang and
457 Song, 2006).

458 It is debated whether Tibet's surface rifting extends through the whole lithosphere
459 (Yin, 2000) or is limited to the upper crust (e.g., McCaffrey and Nabelek, 1998). Our
460 upper-mantle low velocities do not spatially match the surface rifts one to one.
461 Nonetheless, discrete asthenospheric upwellings could create a west-east basal shear
462 on the Tibetan crust that stimulates surface rifting (Tian et al., 2015).

463 The extensional and strike-slip sub-Moho earthquakes (~80-95 km) in southern

464 Tibet (de la Torre et al., 2007; Jiang et al., 2009) may reflect our inferred lithosphere
465 fragmentation. However, if crustal extension is directly related to lithospheric
466 fragmentation, the fragmentation should start at least as early as the surface rifting
467 process, for which the oldest dates yet reported are 19 Ma (Mitsuishi et al., 2012). The
468 earliest west-east extension yet known is a few million years younger than the
469 initiation of the widespread potassic volcanism (~25 Ma) that may alternatively or
470 additionally be related to lithospheric fragmentation.

471

472 **6 Conclusions: An updated three-dimensional model for the Indian lithosphere**

473 Using our updated finite-frequency body-wave tomography results in the
474 south-central Tibetan Plateau, we update our 3D dynamic model of the Indian
475 lithosphere from Liang et al. (2012) (Figure 10). Indian continental lithosphere is
476 currently subducting beneath southern Tibet. This lithosphere can be seen in our
477 model as seismically faster, presumably colder material dipping northward at $\sim 40^\circ$,
478 reaching the BNS at a depth of 350 km. Beneath southern Tibet we observe
479 alternating low- and high-velocity anomalies from west to east that we believe
480 indicate lateral fragmentation of the ICL perpendicular to the India-Asia convergence
481 direction between India and Eurasia. At the same time gravitational detachment of the
482 subducted ICL may be developing along the strike of the collisional mantle front from
483 east to west. We infer that fragmentation has induced local upwelling of
484 asthenosphere beneath the subducting ICL, imaged as low-velocity anomalies that
485 separate the strong fast-velocity anomalies of the subducting ICL. The most

486 significant low-velocity anomaly, i.e. our largest proposed tear, is west of the YGR,
487 extending far to the north and connecting with the low-velocity anomaly beneath the
488 Qiangtang Terrane. Indian asthenosphere may be ascending west of the YGR and
489 flowing northward to mix with asthenosphere beneath northern Tibet.

490 Establishing 3D images of the subducting ICL is an essential step towards a full
491 understanding of the Indian-Eurasian continental collision and the construction of the
492 Tibetan Plateau. Detailed surface-wave tomography using similar data sets will
493 provide a solid test of our 3D model of the upper-mantle structure of the south-central
494 Tibetan Plateau, while joint inversion of body- and surface-wave tomography will
495 further improve the details of the 3D velocity architecture.

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518 **Figure Caption**

519 **Figure 1.** Tectonic and topographic map of the south-central Tibetan Plateau and
520 seismic stations used in this study (XF: Hi-CLIMB, XE: Namche Barwa, YL:
521 HIMNT, XA: Bhutan, XR: INDEPTH II&III, RS: RISE, CD: TIBET-31N).
522 Blue-shaded region: high-velocity anomaly from Tilmann et al. (2003) P-wave
523 tomography. Blue, purple, and yellow lines: 100, 200, and 300 km depth contours on
524 top of Indian lithosphere from Li et al. (2008) P-wave tomography. Structures from
525 the HimaTibetMap dataset (Styron et al., 2010) include MFT: Main Frontal Thrust,
526 IYS: Indus-Yarlung Suture, BNS: Bangong-Nujiang Suture, JS: Jinsha Suture, JF:
527 Jiali strike-slip fault, GF: Gyiring Co strike-slip fault, CR: Cona Rift, YGR:
528 Yadong-Gulu Rift, PXR: Pumqu-Xianza Rift and TYR: Tangra Yum Co Rift.
529 Grey-shaded areas on the Indian plate: Faizabad and Munger-Saharsa basement ridges
530 (Godin and Harris, 2014). Upper-right inset shows events used in this study. Black
531 and red dots show event locations for P- and S-wave event locations, respectively.

532

533 **Figure 2.** Sector diagrams of azimuthally averaged differential travel times for
534 TIBET-31N, for P (a and c) and S (b and d), before (a and b) and after (c and d) the
535 tomographic inversion, respectively. Travel-time residuals before the inversion
536 represent the differential travel times from MCCC measurements. Travel-time
537 residuals after the tomographic inversion represent the difference between the
538 measured differential travel times and the synthetic travel times calculated based on
539 inverted V_p and V_s models. Sector lengths show normalized measurements in a 30°

back-azimuth range at each station. The color of each 30° back-azimuth is based on the averaged travel-time residuals in this back-azimuth range. Because of the large number of data, the averaged travel-time residuals reflect the general travel-time residual values in that back-azimuth range. These travel-time residuals average different ray-paths with similar back-azimuths that have sampled similar regions beneath our study area.

546

Figure 3. Histograms of differential travel times from MCCC measurements (δT_{P_real} , δT_{S_real}), synthetic travel times from final tomographic models (δT_{P_syn} , δT_{S_syn}) and their differences ($\delta T_{P_syn} - \delta T_{P_real}$, $\delta T_{S_syn} - \delta T_{S_real}$) for both P- and S-waves. The black bars show the histogram of both high- and low-frequency data, blue bars show the histogram of high-frequency data, and red bars show the histogram of low-frequency data. The mean values and standard deviations of the different data group are shown using the same color as the bars. The width of each bar is 0.1 second.

554

Figure 4. Three-dimensional checkerboard test results using the same damping parameters as our final models. Checkers have a width (half-wavelength) of ~170 km. Blue and red dashed lines overlain on the output images are $\pm 2\%$ for the P-velocity ($\pm 4\%$ for S-velocity) contours of the starting model. Open inverted triangles show the intersection points of south-north and west-east cross-sections.

560

Figure 5. P- and S-model recovery tests. (a) The input model has a high-velocity slab

562 with a moderate subduction angle and an uppermost-mantle low-velocity anomaly. **(b)**
563 The input model only has a high-velocity slab. The top panels show the input models.
564 The inverted models demonstrate good recovery of both high- and low-velocity
565 anomalies.

566

567 **Figure 6.** Velocity structures beneath southern and eastern Tibet from our
568 finite-frequency teleseismic travel-time tomography. Horizontal slices of P- and
569 S-wave velocities are shown at depths of 84 km, 169 km, and 253 km. Vp and Vs
570 images are clipped based on data coverage (the intensity for P- and S-wave
571 travel-time sensitivity kernels, Figures S1 and S2). The geological structures of Figure
572 1 are shown as grey lines. ICL: Indian Continental Lithosphere; HV or HVBNL:
573 High-Velocity anomalies Beneath the Northern Lhasa Terrane; LV or LVBI:
574 Low-Velocity anomalies Beneath the IYS; LVBNT: Low-Velocity Beneath Northern
575 Tibet.

576

577 **Figure 7. (a)** Five south-north and **(b)** three west-east vertical cross-sections through
578 our P- and S-wave velocity models. Topography on the vertical sections is
579 exaggerated x10. The top 50 km of the model has been removed because crustal
580 structure is poorly constrained. Arrows above each section mark the IYS and BNS.
581 Open inverted triangles show the intersection points of south-north and west-east
582 cross-sections. Velocity anomalies are labeled as in Figure 6.

583

584 **Figure 8.** 3D iso-surface plots for the S-wave model. **(a)** A west-east view and **(b)** a
585 south-north view of the 3D S-wave model, labeled as in Figure 6. The different blue
586 iso-surfaces are showing 1%, 2%, 3% and 4% high-velocity anomalies, respectively.
587 Low-velocity anomalies are not shown (see Figure S4).

588

589 **Figure 9.** Speculative models for the temporal evolution of upper mantle convergence
590 between India and Tibet. The dashed rectangle encloses the current snapshot for both
591 scenarios based on our tomographic images (Figure 6, B-B'). Dark-green regions are
592 continental mantle lithosphere. **(a)** Left panel: underthrusting Indian continental
593 lithosphere rolls back to become a subducting slab. **(b)** Right panel: underthrusting
594 Indian continental lithosphere is thickened during collision then later lost due to
595 Rayleigh-Taylor instability. Each small square on the third plot (current snapshot)
596 represents an area of 400 km².

597

598 **Figure 10.** Cartoon interpretation of the current upper-mantle structure beneath
599 south-central Tibet. The fragmented ICL is helping to drive convection beneath
600 northern Tibet. The lithospheric mantle of northern Tibet is heated by both viscous
601 strain heating and small-scale mantle convection induced by the downwelling of the
602 ICL. Upwelling asthenosphere through gaps in the ICL flows laterally above the
603 subducting slab. Black arrows show the direction of mantle material flow.

604

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Figure 1

Figure 2

Figure 3

Figure 4

Figure 5

Figure 6

Figure 7a

Figure 7b

Figure 8

Figure 9

Figure 10