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Mantle dynamics beneath the Sichuan Basin and Eastern Tibet from teleseismic tomography

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Key Points:

- A new seismic velocity model that constrains the uppermost 700 km of mantle structure around the Sichuan Basin
- Yangtze subduction has weakened the lithosphere to the east of the Sichuan Basin facilitating surface uplift since the Cenozoic
- Lithospheric delamination in eastern Tibet promotes the formation of the Xianshuihe fault and the eastward extrusion of Tibet

Abstract

The cratonic Sichuan Basin is located east of the Tibetan Plateau, and is surrounded by mountains that have undergone complex deformation and uplift since the Cenozoic. Imaging mantle structure is important for understanding its formation, and to date most models suggest a deep cratonic root underlies the basin, blocking eastward extrusion of

lithospheric material beneath Tibet. Here we obtain detailed upper mantle structure from teleseismic tomography in the region utilizing travel time data from earthquakes recorded at 506 seismic stations, including 25 new stations in the poorly sampled Sichuan Basin. Contrasting to previous models, we show eastward and southeastward dipping high-velocity anomalies extending eastward ~150 to 400 km into the upper mantle from the Sichuan Basin. We suggest, the southeastward subduction of the Yangtze Block occurred in the Mesozoic and may be reactivated in the Cenozoic, with the relatively thin and weak lithosphere to the east of the Sichuan Basin prone to deformation in response to the eastward growth of the Tibetan Plateau. A west-dipping high-velocity anomaly beneath eastern Tibet is interpreted as delaminated lithosphere. This delamination may accelerate the development of the Xianshuihe fault zone and the horizontal extrusion of the Tibetan Plateau. Beneath the East Qinling orogen, the eastward extrusion of the plateau material is not obvious suggesting limited horizontal lithospheric extrusion is present north of the Sichuan Basin.

1. Introduction

The Sichuan Basin is an example of a compressive tectonic basin (Figure 1), located in the western Yangtze Block (South China) east of the Tibetan Plateau. The basin is surrounded by orogenic belts which have undergone different tectonic deformation (Shen, Mei, Xu, & Tang, 2007). Eastern Tibet, with relatively thick crust and thin lithosphere, is characterized by high topography and strong crustal deformation, accompanied by a present state of uplift (J. Hu et al., 2015; Liang et al., 2013; Z. Zhang et al., 2010).

Cenozoic surface uplift occurs in the eastern Sichuan basin and further east, the Eastern Sichuan Fold Belt has been further uplifted since 12 Ma (S. Li et al., 2011). GPS results show that these regions to the east of the Sichuan Basin, especially the Hubei-Hunan-Guizhou fold belt, are still being uplifted today (M. He et al., 2018). This makes the region an excellent natural laboratory to study the formation of compressional basins and understand the extrusive tectonics of the eastern Himalaya. However, there are at present few studies focused on the Cenozoic uplift mechanism in the southern and eastern areas of the Sichuan Basin, and the discussions have been mainly limited to the genesis of compression (S. Li et al., 2011; Shen, Mei, Xu, Tang, & Tian, 2007; H. Shi et al., 2016). The study of the effects of deep dynamic processes on surface uplift is thus poorly understood.

Subduction and delamination are principal mechanisms for recycling continental and oceanic lithosphere into the deep mantle (Stern, 2002; Z. Wang et al., 2018; Y. Zheng & Chen, 2016). Oceanic subduction is a well-understood process (e.g. Stern, 2002). On the other hand, there is less attention made to continental subduction. Subduction of the continental lithosphere was long considered impossible due to the buoyancy of continental rocks (McKenzie, 1969), but it is now widely accepted (Stern, 2004; W. Sun et al., 2019; Y. Zheng & Chen, 2016). The subduction of a continental block beneath another continental block results in mountain ranges such as the Alps and Himalayas (Y. Zheng & Chen, 2016). As a response of India-Eurasia continental collision and Indian subduction, horizontal compression is transmitted to regions further north, and an old

69 orogenic belt (Tien Shan) is rejuvenated (Molnar & Tapponnier, 1975). The
70 Tarim basin may have subducted beneath the Tien Shan under
71 compression (Yu et al., 2017; J. Zhao et al., 2003). Nevertheless, continental
72 subduction is still less common and less studied compared to the oceanic
73 subduction.

74 Lithospheric delamination is an important geodynamic process. It results in
75 thermal upwelling and subsequent lithospheric thinning and rapid uplift of
76 the surface (Kay & Mahlburg Kay, 1993; Nelson, 1992; Seber et al., 1996).
77 Further, heating and weakening of the crust due to mantle upwelling to
78 shallow depths means the crust is prone to deform and can be significantly
79 thickened under strong compression, which further contributes to surface
80 uplift. Delamination has been used to explain crustal thickening and
81 thinning of the lithospheric mantle in orogens including the Central Andes
82 and Himalayas (e.g., Krystopowicz & Currie, 2013), as well as the surface
83 uplift and magmatism in regions such as Colorado plateau, the Tibetan
84 plateau, the Appalachians and the Apennines (e.g., Seber et al., 1996).
85 Numerical simulations show that lithospheric delamination may develop
86 through sublithospheric mantle convection (Morency & Doin, 2004),
87 sublithospheric mantle intrusion by retreat of subducting plate (e.g., Gray
88 & Pysklywec, 2012), or metamorphic eclogitization in lowermost crust
89 during crustal shortening and thickening (e.g., Krystopowicz & Currie,
90 2013).

91 Although the deep structures beneath the Sichuan Basin and surrounding
92 regions have been studied using seismic tomography methods, the results

are significantly different, which limits our understanding of the deep dynamic processes. For example: high-velocity anomalies are featured below the Sichuan Basin in various imaging studies, but their distribution and depth extent are still in debate (French & Romanowicz, 2014; Y. Hu & Wang, 2018; Huang & Zhao, 2006; Lei & Zhao, 2016; C. Li & van der Hilst, 2010; Obrebski et al., 2012; F. Zhang et al., 2018). Similarly, there are also different perspectives on whether a low-velocity flow channel exists in the Qinling-Dabie Orogen, north of the Sichuan Basin (Y. Hu & Wang, 2018; Lei & Zhao, 2016; Obrebski et al., 2012; Wei et al., 2017), thus limiting our understanding of the extent of extrusive flow in the lower crust/upper mantle in this region.

Significant imaging differences are mainly due to the sparse distribution of seismic stations, especially the absence of dense seismic stations inside the Sichuan Basin, resulting in the low resolution of velocity models. In this study, we use both temporary and permanent seismic stations, including 25 newly deployed stations within the Sichuan Basin, combined with an improved teleseismic imaging technique, to invert a higher resolution upper mantle velocity model. Our new results reveal the Yangtze Block subduction, and delamination in the eastern Tibet and no evidence for low velocities to the north linked with extrusive flow.

2. Data and Method

In this study, we collected 82,168 P-wave arrival-time data from seismograms of 747 teleseismic events (30° - 90° , $M \geq 5.5$) recorded by 506

seismic stations (Figure 2), including 184 permanent stations belonging to the Chinese National Seismic Network (2007.12-2009.07, 2013.01-2017.11) (X. Zheng et al., 2010), and 297 broadband temporary stations belonging to the western Sichuan seismic array (2006.10-2009.07) (Q. Liu et al., 2014). The 25 broadband temporary stations (2014.10-2016.12) that we deployed within the Sichuan Basin have a recording period of more than 2 years, with 209 clearly recorded events used in this study. These stations approximately cover the study region uniformly and significantly improve the station density in the Sichuan Basin.

Each of the selected events were recorded by at least 26 stations. There were 236 events recorded by 50-100 stations, 351 events recorded by 101-200 stations, and 85 events recorded by more than 200 stations. The maximum recording for a single event was 351. The selected teleseismic events have a good azimuthal coverage around the study region (Figure 2b), which suggests good crossing rays in the upper mantle of our study region.

The selected teleseismic waveforms were processed to remove the mean and trend, then band-pass filtered (0.02-1.0 Hz). The waveform cross-correlation method was applied to calculate the first arrival P-wave travel time (Rawlinson & Kennett, 2004). In order to avoid the impact of the origin error, the mislocation of the source and the velocity anomaly of the source region, we use the relative travel-time residuals in the tomographic inversion (D. Zhao et al., 1994). The relative travel-time residual is given by using the travel-time residual of each station minus the average travel-time residual of all stations:

$$r_{ij}^{relative} = r_{ij} - \bar{r}_i$$

$$\bar{r}_i = \frac{1}{m_i} \sum_{j=1}^{m_i} r_{ij}$$

$$r_{ij} = T_{ij}^{cal} - T_{ij}^{obs}$$

140 Where $r_{ij}^{relative}$ is the relative travel-time residual of the i th event to the j th
 141 station, $\bar{r}_i$ is the average travel-time residual of the i th event, m_i is the
 142 number of stations recorded to the i th event, r_{ij} is the travel-time residual,
 143 T_{ij}^{obs} is the observed travel time, and T_{ij}^{cal} is the synthetic travel time
 144 calculated by 3-D ray tracing (D. Zhao et al., 1992) under the IASP91 model
 145 (Kennett & Engdahl, 1991).

146 The IASP91 model is an average layered model and does not include the
 147 velocity structure of the surface sediment layer, we extract the thickness
 148 (Figure S1 in the supporting information) and velocity of the sediment layer
 149 from the CRUST 1.0 model (Laske et al., 2012), and combine these with the
 150 IASP91 model. We obtained the ray parameter p for each ray (each
 151 observation data), and calculated the travel time t_{psed} of the ray inside the
 152 sediment layer and IASP91 travel time t_{iasp} within the depth range of the
 153 sediment layer, then used $t_{psed} - t_{iasp}$ to correct the observed travel time.
 154 This method could reduce the influence of thick sediment layers on the
 155 observation data in the Sichuan Basin, thus the synthetic travel time which
 156 contains the velocity and travel time of the ray in the sediment layer could
 157 be more accurate. We also designed several models to test the influence of

crustal thickness errors on the inversion results (Text S1 and Figure S2 in the supporting information).

When using relative travel-time residuals, the imaging results may be adversely affected if the station distribution has strong non-uniformity. In order to reduce the influence of uneven station distribution, we change the method from using the relative travel-time residual with equal weight for each station, to one where we normalise the residuals by the number of stations within a $0.5^\circ \times 0.5^\circ$ grid mesh (Figure 2a). We first calculate the average travel-time residual of stations in each grid ($RESe$), then calculate the average of $RESe$ of all grids ($RESa$). Finally, we subtract $RESa$ from the residual of each station. The final value is the relative residual for each station, namely:

$$r_{ijk}^{relative} = r_{ijk} - \frac{1}{N} \sum_{j=1}^N \left(\frac{1}{n_j} \sum_{k=1}^{n_j} r_{ijk} \right)$$

$$r_{ijk} = T_{ijk}^{cal} - T_{ijk}^{obs}$$

Where $r_{ijk}^{relative}$ is the relative residual of the i th event to the k th station in the j th grid, N is the total number of grids divided by the study region, and n_j is the number of stations in the j th grid, r_{ijk} is the travel-time residual, T_{ijk}^{obs} is the observed travel time, and T_{ijk}^{cal} is the synthetic travel time calculated by 3-D ray tracing (D. Zhao et al., 1992) under the IASP91 model (Kennett & Engdahl, 1991) combined with the sediment layer model. Figure 3a shows the distribution of the travel-time residuals. There are more positive residuals than negative residuals for the raw travel-time residuals

(i.e., the difference between observed and calculated travel time). The relative travel-time residuals are normally distributed. The two-sigma range (95.5%) of relative residuals is from -1.92 s to 1.68 s.

In this paper, we applied the teleseismic tomography method of D. Zhao et al. (1992, 1994) to invert the 3-D P-wave velocity structure. Ray tracing is based on a 3-D velocity model during inversion. The 3-D velocity model is composed of an initial velocity model and a 3-D perturbed velocity model. The initial velocity model consists of three layers that are separated by the Conrad and Moho discontinuities. The P-wave velocity between the surface and Conrad is 6.00 km/s, and between Conrad and Moho is 6.67 km/s. IASP91 model is applied for the velocities beneath Moho. The Moho geometry was referenced to our crustal thickness model which was estimated from receiver functions (W. Wang et al., 2017a; W. Wang et al., 2017b) (Figure S1 in the supporting information). Given a point in the model space, we can obtain the corresponding depths of Conrad and Moho, and determine which layer this point is located in, then calculate the initial velocity. The 3-D perturbed velocity model is represented by velocity perturbations of grid nodes. The velocity perturbation of any point in model space is obtained from linearly interpolating the velocity perturbations of its surrounding 8 nodes. The velocity could be calculated from the initial velocity and the velocity perturbation. A pseudo-bending method (Thurber & Um, 1987) and Snell's law are used to perturb the ray path for ray tracing in continuous space and at the discontinuities, respectively, from which travel times could be obtained (D. Zhao et al.,

1992). A new 3-D velocity model is obtained after each iteration and used for ray tracing and travel time calculation in the next iteration. The LSQR algorithm was used to solve the large observation matrix (Paige & Saunders, 1982).

The area investigated lies between latitude 26° and 35° north and longitude 100° and 112° east. In order to parameterize the study region, a model of $0.5^{\circ} \times 0.5^{\circ}$ in the lateral direction and -200, 0, 50, 100, 200, 300, 400, 500, 600, 700, 800, 900 km in the depth direction was set. Note that the -200, 800 and 900 km are only used for interpolation and are not reflected in the final imaging results. In order to avoid the influence of possible errors at the edge of the model, the actual model extends outwards along the latitude and longitude directions during parameterization, namely across latitude range of 22°N - 39°N and longitude range of 96°E - 116°E , and the node spacing is still 0.5° . In the inversion process, the damping factors were introduced to suppress sudden changes in small scale structure. The damping factor was finally selected as 7 after repeated tests (Figure 3b).

Due to the non-uniformity of the epicenter distribution, the final imaging results may produce tensile distortion along the direction of densely distributed epicenters. We improve the inversion method by the addition of gridding and earthquake weighting to the original program. We divide all the regions containing the teleseismic epicenter into a grid mesh of $5^{\circ} \times 5^{\circ}$. If the number of earthquakes within a single grid is less or equal than 3, the weight of each earthquake within the grid is the number of earthquakes divided by 3. Otherwise the weight is 1 divided by the number of

earthquakes. This method reduces the adverse effects of excessive epicenter distribution in some regions on the imaging results.

The parameterized model in the study region (100°-112°E, 26°-35°N, 0-700 km) contains a total of 4,275 grid nodes. There are 4,230 grid nodes in the model with hit counts (number of ray paths passing around each grid) exceeding 10, and 4,080 nodes exceeding 100 after inversion, indicating that most nodes have a sufficient number of crossing rays to ensure good ray coverage in the study region.

In order to further evaluate the reliability of the inversion results, we conducted checkerboard resolution tests. Alternate $\pm 2.5\%$ velocity perturbations were given to adjacent nodes. Synthetic travel times are calculated with the addition of random noise of standard deviation of 0.1 s applied. The synthetic travel time data set was calculated using the same event-station paths as the actual observation data, and then the velocity structure was inverted. The resolution of the study region is visually evaluated by the inversion results of the checkerboard model. Figure 4 shows the test results at different depths. The velocity values are well recovered at each depth, with a lateral resolution in most regions approaching 0.5°. However, due to the limitation of station distribution in some regions where the station spacing reaches ~100-150 km, the resolution may be lower than 0.5° at shallow depths, but could reach 0.8° below 100 km in these areas (Figure S3 in the supporting information). According to the checkerboard results, the 25 temporary stations which are deployed within the Sichuan Basin have significantly improved the

resolution of the region, especially at 50 km depth. This is particularly noticeable when comparing resolution in this region with the southeast side of the Sichuan Basin. We also conducted spike tests since discrete spike tests are more useful for assessing the resolving power of the data set to recover structure compared to traditional checkerboards (Rawlinson & Spakman, 2016). The results show that the resolution is relatively low in regions with sparse station density, but the anomalies are largely undistorted in each layer, even at great depths (Figures S4, S5 in the supporting information). This verifies that our improved method is effective.

3. Results and Discussion

Figure 5 and 6 illustrate map views and vertical sections of our P-wave teleseismic tomographic model, respectively. The results of the $0.5^{\circ} \times 0.5^{\circ}$ (Figure 5) and $0.8^{\circ} \times 0.8^{\circ}$ (Figure S3 in the supporting information) grids are essentially the same. In the following discussion we focus on the $0.5^{\circ} \times 0.5^{\circ}$ models. The results of restoring resolution test are shown in Figure 7. The processes are the same as those of the checkerboard resolution test, but the input models are different. We created an input model with fast velocity perturbations of 2% and 0.8% (Figure 7a), and an input model with perturbations of 2%, 0.8% and -2% (Figures 7c). The test shows good recovery though some smearing is apparent below ~ 400 km (Figures 7b, d).

3.1 Deep structure of the Sichuan Basin and surrounding regions

273 Relatively high-velocity anomalies are located beneath the Sichuan Basin,
274 and large-scale low-velocity anomalies are detected beneath the eastern
275 Tibetan Plateau, west of the Sichuan Basin (Figure 5a-d). The sharp velocity
276 contrast along the Longmenshan fault zone is consistent with the overall
277 characteristics of previous results (Y. Hu & Wang, 2018; Lei & Zhao, 2016;
278 Obrebski et al., 2012; Wei et al., 2017), but resolution is improved in our
279 models by the addition of temporary seismic stations within the Sichuan
280 Basin. The resolution inside the basin could reach 0.5° at various depths
281 (Figure 4). In cross-sections A-A' and B-B' (Figure 6), the velocity contrast
282 can be traced down to a lithospheric-scale depth of ~ 150 km beneath the
283 Longmenshan fault zone.

284 Velocity profiles across the Longmenshan fault zone reveal that the
285 distribution of crust and upper mantle velocities are well correlated with
286 the surface topography (Figure 6A, B). The low-velocity zone underlying
287 high surface topography west of the Longmenshan fault zone is in sharp
288 contrast with the high velocity underlying low topography in the Sichuan
289 Basin. The large scale low-velocity anomaly in the lower crust and upper
290 mantle is commonly associated with the presence of warm materials.
291 Compared with the Sichuan Basin, the upper mantle of the Tibetan Plateau
292 has higher temperature (Y. Sun et al., 2013) and thinner equivalent elastic
293 thickness of the lithosphere (Jordan & Watts, 2005), which is consistent with
294 our result. The relatively high temperatures may lead to a decrease in the
295 crust and upper mantle strength of eastern Tibet, and make the crust and
296 upper mantle structure easy to deform under the effect of extrusion of the

Tibetan plateau, resulting in crustal thickening and surface uplift. The high-velocity upper mantle materials in the Sichuan Basin are characterized by low temperature (Y. Sun et al., 2013), high density (Lou et al., 2008), high strength and weak deformation (Gao et al, 2019; X. Wang et al., 2018), and thus have a strong resistive effect in the process of eastward extrusion of upper mantle materials from the Tibetan Plateau (Clark & Royden, 2000).

The low-velocity anomaly beneath eastern Tibet extends into the Sichuan Basin (Figures 5c-e, Figures 6A-B). The Longmenshan fault zone located in the northwest of the Sichuan Basin is today characterized by intense seismic activity. Shear wave splitting observations (Gao et al., 2019) show that the fast polarization directions near the Longmenshan fault zone are oblique to the strike of the faults, and the authors inferred that the western margin of the Sichuan Basin may have been modified. The warm low-velocity anomaly passes through the Longmenshan fault zone and extends beneath the high-velocity lithosphere of the western Sichuan Basin, which may indicate that the eastward expansion of the Tibetan Plateau is reactivating the lithosphere of the western Sichuan Basin.

Previous seismic observations revealed clear upper mantle high-velocity anomalies beneath the Sichuan Basin, however, controversies remain in the distribution pattern and depth extent of high-velocity anomalies. Lei and Zhao (2016) used teleseismic P-wave tomography to illustrate high-velocity anomalies down to ~200 km depth under the Sichuan Basin. Huang and Zhao (2006), C. Li and van der Hilst (2010) and F. Zhang et al. (2018)

displayed high-velocity anomalies extending down to ~300 km depth. Y. Hu and Wang (2018) showed that high-velocity anomalies reached ~400 km. Joint inversion of surface and body waves by Obrebski et al. (2012) showed that high-velocity anomalies reached the bottom of the mantle transition zone, about 700 km depth. The lack of seismic stations in the Sichuan Basin may be responsible for the different results obtained by various researchers. Despite these differences, most studies suggest that fast velocities in the mantle beneath the Sichuan Basin show a deep cratonic root is present, a remnant of the South China Block (e.g., Y. Hu & Wang, 2018; Obrebski et al., 2012; F. Zhang et al., 2018).

Our new imaging results, using the dense station distribution in the Sichuan Basin (Figure 2a) reveal an eastward deepening of the high-velocity bottom. The high velocities extend to a depth of ~150 km beneath the Longmenshan fault zone, gradually deepening from west to east, reaching 250 km in the central basin to more than 400 km at ~110°E (Figures 6A-D, 8a, b and d). The shallower bottom depth in west of the basin represents relatively thin lithosphere compared to that in the central area.

Some researchers suggest that the Yangtze Block collided with the Cathaysia Block during the Mesozoic (Hsü et al., 1988; Hsü et al., 1990; Y. Wang et al., 2003). The location of the western boundary between the Yangtze Block and the Cathaysia Block remains controversial (Zhu et al., 2019, and references therein). Geochemical and receiver function studies suggest that the Xuefeng Mountain area may be the western boundary (Y. Wang et al., 2003; Y. Zhang et al., 2018). In this location, the crust and

345 lithosphere structure on the opposite sides of the mountain have sharp
346 variations (L. Liu et al., 2012; Y. Zhang et al., 2018). The Moho interface dips
347 to the southeast on the west of Xuefeng Mountain, and the Moho depth is
348 marked by an abrupt change of 5-10 km in the Xuefeng Mountain (deeper
349 in the west and shallower in the east) (C. He et al., 2013; S. Li et al., 2019).
350 The composition of Mesozoic granites in South China indicates that the
351 crust of the Yangtze were underthrust beneath the Cathaysia Block in the
352 Mesozoic (Hsü et al., 1988; Y. Wang et al., 2003).

353 Our tomographic images show the eastward dipping high-velocity
354 anomalies reach Xuefeng Mountain area, where the thickness of the
355 anomalies is about 150 km (Figure 6C, G, and 8). Lithospheric thrust faults
356 are distributed to the west of Xuefeng Mountain (S. Li et al., 2019), and their
357 location in the uppermost lithosphere is consistent with the dips present in
358 the high-velocity anomalies. We speculate that the Yangtze Block has
359 subducted beneath the Cathaysia Block in the Mesozoic.

360 A gap is located between the position where the high-velocity anomalies
361 begin to dip and the boundary of the Yangtze Block and the Cathaysia
362 Block, but this is not a rare occurrence. A representative example is
363 India-Eurasia collision (Stern, 2004; Xiao et al., 2017; W. Zhao et al., 1993).
364 The Yarlung Tsangpo suture zone marks the boundary between India and
365 Eurasia. Deep seismic reflection data indicate that the Indian plate
366 subducted northward at the Main Boundary Thrust (MBT), which is about
367 300 km away from the Yarlung Tsangpo suture zone (W. Zhao et al., 1993),

368 and the subduction sites both on the surface and the Moho interface may
369 migrate southward in the future.

370 The structural style west of the Xuefeng Mountain is characterized by a
371 northwestward propagation of thrust-detachment in the early Cretaceous (Y.
372 Wang et al., 2013). This suggests that the subduction site of the Yangtze
373 Block may migrate to the northwest relative to the Xuefeng Mountain area.
374 The relatively thin lithospheric thickness (Y. Zhang et al., 2018) and the
375 large-scale Mesozoic magmatism (S. Li et al., 2019; L. Liu et al., 2012; Y.
376 Wang et al., 2003) to the east of Xuefeng Mountain suggest that the old, cold
377 subducted lithosphere in this area may have been heated by thermal
378 upwelling and thus, the velocity anomaly may no longer be present.

379 Old orogenic belts or suture zones may be reactivated by new tectonic
380 events, leading to intracontinental subduction. For example, the
381 India-Eurasia collision reactivated the old orogenic belt (Tien Shan) 1,000
382 km north of the collision boundary (Molnar & Tapponnier, 1975).
383 Magnetotelluric sounding profile and 2-D velocity structure show a
384 slab-like structure dipping to the north, suggesting that the lower crust and
385 upper mantle of the Tarim basin has subducted into the upper mantle
386 beneath the Tien Shan (J. Zhao et al., 2003). For our study region, the area
387 between the Sichuan Basin and Xuefeng Mountain was not uplifted during
388 ~35-15 Ma, with uplift occurring post 15 Ma (H. Shi et al., 2016; Tang et al.,
389 2014). GPS results show that the crust of the South China Block is
390 characterized by eastward movement (J. Wang et al., 2008). The weak
391 lithosphere of the eastern Yangtze Block, which manifests itself as a low

seismic velocity compared to the Sichuan Basin, may be inherited from the complex tectonic history, such as the collision between the Yangtze and Cathaysia blocks and the hydration related to the oceanic subduction before the collision (Hsü et al., 1990; Y. Wang et al., 2003), and it may be undergoing strain concentration that is prone to be reactivated. However, we cannot wholly exclude the possibility that the subduction of the Yangtze Block has been reactivated due to the Cenozoic eastward extrusion of the Tibetan Plateau. There are two NE-SW weak seismic belts in eastern Sichuan Basin from seismic data (Figure 9), and the location of the belts is roughly consistent with the location of the eastward dipping high-velocity anomaly at shallow depth. A west-east profile at 30°N, shows an eastward dipping seismicity at ~108°E (Figure 9), which may be an indicator of subduction reactivation. However, focal depths may be poorly constrained relative to epicenter locations especially for events before the year 2007 and so this needs further research. The rigid Sichuan Basin may transmit the eastward compressive stress of the Tibetan Plateau to the Eastern Sichuan fold belt and the Hubei-Hunan-Guizhou fold belt, leading to the Cenozoic uplift of these areas (1-2 mm/yr) (M. He et al., 2018; H. Shi et al., 2016; Yuan et al., 2018), similar to the way that the Tarim Basin transmits compressive stress to the Tien Shan (Molnar & Tapponnier, 1975), leading to the uplift of the orogenic belt.

The Qinling orogenic belt and Daba Mountain area, the boundary of the Mesozoic plate collision zone located north of the Sichuan Basin, are characterized by relatively low seismic velocity (and thus low strength)

relative to the Sichuan and Ordos basins (e.g. Y. Hu & Wang, 2018). Daba Mountain has been uplifted at 120-110 Ma (Shen, Mei, Xu, Tang, et al., 2007), which indicates that Daba Mountain has been relatively weak in the Mesozoic. The uplift and eastward extrusion of the Tibetan Plateau have produced eastward compression on the Sichuan Basin and Qinling orogenic belt since the Cenozoic (H. Shi et al., 2016; X. Shi et al., 2020). The high-velocity lithosphere in the Sichuan Basin represents relatively high strength (Jordan & Watts, 2005), so no significant deformation occurred (J. Liu et al., 2020). The Qinling orogenic belt and Daba Mountain area are prone to deform due to relatively low strength of lithosphere. The Qinling orogenic belt show a gradual decrease in elevation from the west to the east (Figures 1 and 6E), indicating that the deformation in the west is larger than the east. The deformation and uplift of Daba Mountain is controlled by the E-W compressive stress (X. Shi et al., 2020).

3.2 Lithospheric delamination in the eastern Tibet

Our tomographic images highlight a progressive westward dipping high-velocity anomaly underlying the low-velocity anomalies located in eastern Tibet (Figures 5c-e, 6B, C, 8a-d). This extends from the bottom of the lithosphere beneath the Sichuan Basin to the 410 km discontinuity (~150-450 km), and the thickness increases with depth to about 150-km thick at the bottom. We interpret the high-velocity body as delaminated 150 km thick lithospheric material. Receiver function imaging results of J. Hu et al. (2015) show that the 410 km discontinuity is uplifted and the 660 km discontinuity is depressed along 30°N in the eastern Tibet. The distribution

of this uplifted area is consistent with the lithosphere delamination in our imaging result, explained by a lateral decrease in temperature at the level of the transition zone uplifting the 410 km discontinuity (Bina & Helffrich, 1994). Chen et al. (2014) found weak rheological strength of the lithospheric mantle at this location, which supports delamination beneath eastern Tibet. An excess pressure brought by rising asthenosphere after delamination is then responsible for the growth of the eastern Tibetan margin.

Similar to our models, previous studies have revealed a significantly thinned lithosphere beneath the eastern Tibetan Plateau compared with a thicker lithosphere beneath the Sichuan Basin (J. Hu et al., 2015; Lou et al., 2008; Z. Zhang et al., 2010). We speculate that the relatively thick lithosphere beneath the eastern Tibetan Plateau near 30°N may have been heated and weakened by the high-temperature, low-velocity upper mantle beneath the Tibetan Plateau, and then thickened under strong compression. With the further increase in temperature of the thickened lithosphere, the strength of lithosphere decreases, causing the relatively cold, dense lithosphere (compared with upper mantle materials of Tibetan Plateau) to delaminate.

Gongga Mountain, the highest peak (7556 m) in the eastern margin of the Tibetan Plateau, is located above the delaminated lithosphere revealed in this paper (Figures 5c-e, 8a, c and d), and is still uplifting. Its current uplift rate is 1-3 mm/yr (Liang et al., 2013; Pan & Shen, 2017). The Gongga Mountain area has high heat flow and many hot springs (Hochstein &

Regenauer-Lieb, 1998) and consists of two rock units: The Gongga granite pluton and an elongate migmatite zone (H. Li & Zhang, 2013). SHRIMP U–Pb isotopic dating of samples from this region yielded the Miocene granite emplacement (18–12 Ma), and the possible existence of crustal component mixed with mantle-derived magmas (H. Li & Zhang, 2013; Roger et al., 1995). Kay and Mahlburg Kay (1993) regarded magmas as containing the best evidence for recognizing delamination events. The partial melting materials in this region have undergone high-pressure metamorphism (46–18 Ma) after strong compression, and there are garnet-rich metamorphic rocks in the source rocks, while partial melting residuals dominate garnet, and the melting formation environment is similar to adakite (S. Liu et al., 2006). The density increase caused by the eclogite in the crust could promote the development of thick lithosphere in this area which is then prone to delaminate (Krystopowicz & Currie, 2013). Therefore, the gravity instability caused by dense garnet may lead the lithosphere to delaminate. We speculate that the formation of high-altitude peaks in Gongga Mountain and surrounding areas, and the Miocene granite emplacement, are closely related to the lithospheric delamination and the subsequent eastward extrusion of Tibetan Plateau.

The NW–SE striking Xianshuihe fault zone is an important active lithospheric scale strike-slip fault in the eastern Tibetan Plateau (Roger et al., 1995), with a prominent bend near Gongga Mountain (Cook et al., 2018). GPS observation results indicate the current slip rate of ~10–12 mm/yr (E. Wang & Burchfiel, 2000; P. Zhang et al., 2004), and left-lateral shearing of the

southeastern segment has occurred since at least 18-12 Ma (H. Li & Zhang, 2013; Roger et al., 1995), quite similar to the age of emplacement of the Gongga granite, and these ages correspond with the possible time required for the delaminated lithosphere to reach the transition-zone depths (~20 Ma) (Schott & Schmeling, 1998). Along the eastern segment of the Xianshuihe fault zone there is evidence of high temperature metamorphism during the early Cenozoic (H. Li & Zhang, 2013), which may be due to the upwelling of hot materials after delamination heating the crust. The formation of the fault allowed the extrusion of the Tibetan Plateau toward the southeast (Roger et al., 1995). We suggest that the lithospheric delamination adjacent to Gongga Mountain provided the heat needed to weaken the crust, which accelerated the formation of the Xianshuihe fault zone and promoted the extrusion of the Tibetan Plateau.

3.3 Velocity structure of Qinling-Dabie orogen

Low-velocity anomalies are present beneath eastern Tibet, widely distributed beneath the West Qinling orogen, Songpan-Ganzi block and Chuan-Dian diamond block, and extend down to ~500 km depth (Figures 5a-f and 8c). Previous studies also observed large-scale low-velocity anomalies west of the Longmenshan fault zone, however, it is still controversial whether an obvious low-velocity anomaly, signifying an eastward mantle flow channel, exists beneath the Qinling orogen (Y. Hu & Wang, 2018; Lei & Zhao, 2016; Obrebski et al., 2012; Wei et al., 2017). Our results show that no obvious low-velocity anomaly exists in the East Qinling orogen above 100 km depth, rather high-velocity anomalies are present

(Figures 5a, b, 6E, F and 8b, c). This is especially clear in the depth slice at 50-100 km with a resolution of 0.8° (Figure S3 in the supporting information). There are only weak low-velocity anomalies below 100 km depth beneath the Qinling orogen between the Ordos and Sichuan blocks (Figures 5c, 6E, F and 8b). Results from body wave imaging (F. Zhang et al., 2018; L. Zhao et al., 2012), Rayleigh and S wave imaging (Jiang et al., 2013; Wei et al., 2017) and joint inversion of receiver function and Rayleigh wave (X. Wang et al., 2017) show similar patterns. Our results suggest that the eastward extrusion of the plateau material is not obvious north of the Sichuan Basin due to the lack of low-velocity anomaly in the East Qinling Orogen.

4. Conclusions

High-resolution velocity structure of the upper mantle beneath the Sichuan Basin and surrounding regions was obtained through teleseismic P-wave tomography, using 82,168 waveform data from 747 teleseismic events recorded by 506 stations in the study region.

Our results reveal eastward and southeastward dipping high-velocity anomalies in the upper mantle beneath the Sichuan Basin and the Yangtze Block, which extend down to depths of ~ 150 km beneath the Longmenshan fault zone, ~ 250 km in the central basin, and more than 400 km to the east of the basin. These high-velocity anomalies may indicate that the Yangtze Block has subducted beneath the Cathaysia Block in the Mesozoic, and the subduction site may migrate to the northwest. The subduction produced a weak lithosphere, leading to the uplift of the Hubei-Hunan-Guizhou fold

belt and Eastern Sichuan fold belt in response to the eastward growth of the Tibetan Plateau.

A westward dipping high-velocity anomaly exists within the generally low-velocity upper mantle in eastern Tibetan Plateau, extending roughly from the base of the lithosphere beneath the Sichuan Basin to the top of the 410 km discontinuity, with a thickness of ~150 km at the bottom. Taking into account the adjacent high-altitude peaks including Gongga Mountain and the Miocene granite emplacement, we interpret the high-velocity body as delaminated lithosphere. The lithospheric delamination served to heat and weaken the crust, which drives mountain building, accelerates the formation of the Xianshuihe fault zone and promotes the extrusion of the Tibetan Plateau.

No obvious low-velocity anomaly exists in the East Qinling orogen above 100 km depth, rather high-velocity anomalies are present. There are only weak low-velocity anomalies below the depth of 100 km. We therefore suggest that the eastward extrusion of the plateau material is not obvious under the Eastern Qinling orogen.

Acknowledgments

We greatly appreciate editors and anonymous reviewers for their valuable and constructive criticism that improves the presentation of this manuscript. We acknowledge the Data Management Center of China Seismic Array (<http://dx.doi.org/10.12001/ChinArray.Data>), and the Data Management Centre of China National Seismic Network at Institute of Geophysics, China

559 Earthquake Administration (SEISDMC,
560 <http://dx.doi.org/10.11998/SeisDmc/SN>) for providing seismic waveform
561 data. The waveform data used in this study can be downloaded via
562 <https://doi.org/10.6084/m9.figshare.13076750.v1>. The earthquake catalogs
563 were provided by China Earthquake Network Center (CENC,
564 <http://www.ceic.ac.cn/history>). We thank D. Zhao for offering the source
565 codes of travel time tomography. This work was supported by the National
566 Natural Science Foundation of China (Grant Nos. 41774102 and 41374097).

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828 **Figure Captions**

829 **Figure 1.** Regional tectonic and topographic map of the study area (red
830 box). Solid lines represent the main tectonic boundaries and sutures. The
831 blue dashed line represents the North-South Gravity Lineament (NSGL) (Y.
832 Zhang et al., 2018). The yellow lines represent the boundary between the
833 Yangtze Block and the Cathaysia Block. The thick black lines represent the
834 main faults east of the Sichuan Basin. The red triangle represents the
835 Gongga Mountain. Dots denote events (1970-2020, $3 < M_s < 8$) with
836 hypocenter depths of 0-33 (purple), 33-70 (blue), and 70-300 km (green).

837

838 **Figure 2.** (a) The distribution of 506 seismic stations used in this study. The
839 gray inverted triangles show 184 permanent stations, the gray squares
840 show 297 temporary stations of the western Sichuan seismic array (Q. Liu et
841 al., 2014), and the black triangles show 25 broadband temporary stations
842 deployed in the Sichuan Basin (2014.10-2016.12). The grid spacing is
843 $0.5^\circ \times 0.5^\circ$. Other labels are the same as in Figure 1. (b) Epicenter
844 distribution of the 747 teleseismic events (black dots) used in this study.
845 The concentric circles in the insert map correspond to epicentral distance
846 of 30° , 60° , and 90° , respectively.

847

848 **Figure 3.** (a) Histogram of the raw and relative travel-time residuals. (b)
849 Trade-off curve for the norm of solution and RMS (root-mean-square) travel
850 time residual. The damping factor chosen in this study is 7.

851

852 **Figure 4.** Results of checkerboard resolution test for $0.5^{\circ} \times 0.5^{\circ}$ input
853 velocity perturbations, where adjacent nodes are $\pm 2.5\%$, and a random
854 error with a standard deviation of 0.1 s is added. The depth of each layer is
855 shown at the top left of each map. The solid and open circles represent high
856 and low velocities, respectively, and the scale of the velocity perturbation is
857 shown at the bottom. Other labels are the same as in Figure 1.

858

859 **Figure 5.** P-wave velocity images for a $0.5^{\circ} \times 0.5^{\circ}$ grid on each depth slice
860 (percentage relative to average velocity). The depth of each layer is shown
861 at the top left of each map. Red and blue represent low-velocity and
862 high-velocity anomalies, respectively. The red triangle represents the
863 Gongga Mountain. The scale of the velocity perturbation is shown at the
864 bottom. Other labels are the same as in Figure 1.

865

866 **Figure 6.** Vertical cross sections for the $0.5^{\circ} \times 0.5^{\circ}$ grid P-wave velocity
867 perturbations along the lines shown in the inset map. The surface
868 topography is displayed at the top of each section. The dashed lines denote
869 the 410 and 660 km discontinuities. The scale of the velocity perturbation is
870 shown at the bottom. NSGL, North-South Gravity Lineament; ESFB, Eastern
871 Sichuan Fold Belt; HHGFB, Hubei-Hunan-Guizhou Fold Belt; LMSE,
872 Longmenshan Fault Zone.

873

Figure 7. Results of a restoring resolution test. (a) and (c) show the synthetic models. (b) and (d) show the corresponding inverted models. (a) contains high-velocity bodies with perturbations of 2% (middle) and 0.8% (left and right). (c) contains high-velocity bodies with perturbations of 2% (middle), 0.8% (right) and low-velocity body with perturbation of -2%. The scale of velocity perturbation (in %) is shown at the bottom.

Figure 8. Three-dimensional visualization of P-wave velocity structure in the study region. The +0.8% (blue) and -1.5% (red) velocity isosurfaces in (a)-(c) are rendered from velocity model. The opacity of the partial +0.8% and -1.5% isosurfaces is lowered. Three planes showing variations of P-wave velocity are cut at depths of 0, 410 and 660 km. (a), (b), and (c) are the angles of view from the south, north, and top of the model, respectively. The velocity perturbation scale of the horizontal slices is shown on the right. Thick white dashed line in (c) represents the North-South Gravity Lineament. (d) The inferred dynamic model primarily based on the profile CC' in Figure 8 (green dashed line in (c)). Blue (+0.8%) and red (-0.5%) colors represent cold and hot materials, respectively. NSGL, North-South Gravity Lineament; ESFB, Eastern Sichuan Fold Belt; HHGFB, Hubei-Hunan-Guizhou Fold Belt; LMSF, Longmenshan Fault Zone.

Figure 9. Seismicity of the study area. Blue dots represent the hypocentral locations of earthquakes with magnitude greater than 2.0 from 1970 to 2014.

897 The dots in 30°N profile show seismicity within a width of 1°
898 (29.5°N-30.5°N).

899

Figure 1.

Figure 2.

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