Defining the Limits of Greater India

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Abstract Greater India comprises a part of the Indian plate that subducted under Asia to help form the Tibetan Plateau. Defining the size of the Greater India is thus a key constraint to model the India-Asia collision, growth of the plateau, and the tectonic evolution of the Neo-Tethyan realm. We report Early Cretaceous paleomagnetic data from the central and eastern Tethyan Himalaya that yield paleolatitudes consistent with previous Early Cretaceous paleogeographic reconstructions. These data suggest Greater India extended at least 2,675 ± 720 and 1,950 ± 970 km farther north from the present northern margin of India at 83.6 °E and 92.4 °E at ~130 Ma. An area of ≥ 4.7 × 10⁶ km² of Greater Indian lithosphere was consumed by subduction.

Plain Language Summary Greater India is part of the Indian plate, subsequently subducted under Asia, that helped create the Tibetan Plateau. The amount of Greater Indian crust therefore plays a critical role to address key problems in continental geodynamics. To what extent can continental crust be subducted? How much crust was derived from horizontal shortening of existing crust? How much of Tibet was created by subducted buoyant, continental crust? We provide paleomagnetic evidence that defines the minimum size of Greater India. Our data show that a lithospheric area of ≥ 4.7 × 10⁶ km² was subducted, which supports the notion that the growth of Tibetan Plateau in the Cenozoic occurred by adding buoyant material to its base.

1. Introduction

The Indian subcontinent formed a part of eastern Gondwanaland before rifting away in the Cretaceous (Veevers et al., 1975). Paleogeographic reconstructions based on magnetic anomalies and facies analyses suggest that the northern margin of India extended farther off the Western Australian margin in the past (Powell et al., 1988) but has since been subducted under Asia. This lost landmass is known as Greater India. Knowing how much lithosphere was consumed plays a critical role to explain how the Tibetan Plateau was built in time and space as a consequence of the India-Asia collision. However, estimates for the extent of Greater India are highly uncertain—ranging from several hundreds of kilometers to >2,000 km based on a wide variety of observations (Ali & Aitchison, 2005; Powell et al., 1988; van Hinsbergen et al., 2012, 2018; Yi et al., 2011). For example, studies that correlated the morphology of the western Australian margin with Greater India lead to differences for the extent of eastern Greater India by a factor of two: ~1,800 km according to Powell et al. (1988) and ~950 km in Ali and Aitchison (2005). A more accurate understanding of the original extent of Greater India is therefore a key parameter needed to understand the tectonic evolution of the Tethyan realm and how the Tibetan crust became so thick (Staisch et al., 2016).

The boundary between the Indian and Asian plates is demarcated by Jurassic-Early Cretaceous ophiolites that define the Indus-Yarlung suture zone (Figure 1). Sedimentary and volcanic sequences of the Tethyan-Himalaya belt crop out south of the suture until the high Himalayas. The Tethyan-Himalaya belt is thought to consist of parts of India that were obducted during collision (Yin, 2006). These obducted remnants hold clues for reconstructing the history of Greater India.

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Paleomagnetism plays a unique role in constraining the paleogeography of the Tethyan Himalaya, and by inference, Greater India. In some cases, low-grade metamorphism in the Tethyan Himalaya may have contributed to chemical remagnetization of Cretaceous to early Paleogene rocks (Appel et al., 2012; Huang et al., 2017), which has led to ambiguity surrounding the interpretations of the paleomagnetic data (Patzelt et al., 1996; Yi et al., 2011), and hence of Greater India reconstructions. Paleolatitude estimates from Early Cretaceous rocks within the Tethyan Himalaya range from ~25°S to 55°S (Huang et al., 2015; Klootwijk & Bingham, 1980; Ma et al., 2016; Yang et al., 2015; Zhu & Teng, 1984; supporting information Table S1). Explanations for the wide range in paleolatitude data could come from multiple sources, such as the age of magnetization (primary vs. secondary), where the terrain originated on Greater India, and when the terrain became obducted. Only a minor amount of Early Cretaceous rocks has been studied for paleomagnetism from the Tethyan Himalaya (Huang et al., 2015; Klootwijk & Bingham, 1980; Ma et al., 2016; Yang et al., 2015; Zhu & Teng, 1984; supporting information Table S1). Explanations for the wide range in paleolatitude data could come from multiple sources, such as the age of magnetization (primary vs. secondary), where the terrain originated on Greater India, and when the terrain became obducted. Only a minor amount of Early Cretaceous rocks has been studied for paleomagnetism from the Tethyan Himalaya (Huang et al., 2015; Klootwijk & Bingham, 1980; Ma et al., 2016; Yang et al., 2015), which stretches E-W for over 1,500 km along the Himalayan Arc (Yin, 2006; Figure 1b). The limited data hinder our understanding of fundamental questions at the forefront of continental geodynamics. To what extent can continental crust be subducted? How much crust was derived from horizontal shortening of existing crust? How much of Tibet was created by subducted buoyant, continental crust? To address these questions, we report paleomagnetic data from Early Cretaceous rocks from two localities that are separated by ~9° in longitude along strike of the Tethyan Himalaya. The magnetizations were likely acquired during formation of the rocks and provide a minimum estimate for the extent of Greater India.
2. Materials and Methods

Himalayan geology is divided into four litho-tectonic units from south to north: the sub-Himalaya, Lower Himalaya, Higher Himalaya, and Tethyan Himalaya (Figure 1b). The Tethyan Himalayan sequences (THSs) extend ~150 km N-S and ~1,500 km E-W; they formed in a passive continental margin setting (Aikman et al., 2008). The THS is bounded to the north by the Indus-Yarlung suture zone and to the south by the South Tibetan Detachment (Figure 1a). It contains deformed Proterozoic to Eocene siliciclastic and carbonate sedimentary-metasedimentary rocks as well as Paleozoic to Mesozoic igneous rocks (Yin, 2006). The THS was horizontally shortened at least 70–76% by a predominantly regional fold and thrust belt (Aikman et al., 2008).

The two localities we studied, Cuona and Zhongba, are separated ~960 km in the E-W direction (Figure 1). To the east near Cuona (28.1°N, 92.4°E), we drilled and oriented 258 cores (20 sites) in Lakang Formation neritic marly limestones and 68 cores (eight sites) in mafic lavas and sills intercalated and intruded within the limestones (Figures 1 and S1 and Table S2). Some outcrops have pillow textures, suggesting those lava flows were emplaced in water on the sea floor. Bedding attitudes define a cylindrical, nonplunging fold whose axis trends 73.7° (Figure S1). Undeformed 44-Ma granitoids in the region suggest folding occurred before the Eocene (Aikman et al., 2008). SHRIMP U-Pb dating of the intercalated mafic rocks yielded a magmatic age of 131.1 ± 6.1 Ma (Zhu et al., 2009), compatible with the Early Cretaceous age of the limestone based on paleontologic dating of marine fossils (Zhu et al., 2008).

The western locality (29.7°N, 83.6°E) lies 20-km southwest of Zhongba, where we drilled two sections with one sample per stratigraphic horizon (Figures 1 and S2 and Table S3). At the ~100-m-thick Pianji section, 20 cores were collected from Duobeng Formation cherts and 55 cores from Rilang Formation sandstones; 25 chert and 5 sandstone samples were collected in the ~40-m-thick Duoping section. Radiolaria within the cherts are Barremian in age (Early Cretaceous, 131–125 Ma; Li et al., 2017). These cherts conformably overlie volcaniclastic sandstones in both sampled sections (Figure S2), which have a maximum deposition age of 134 ± 4 Ma constrained by detrital zircon U-Pb ages at the Pianji section (Du et al., 2015).

Samples from Lakang Formation and Rilang-Duobeng Formation were treated in a magnetically shielded room with residual fields less than 500 nT at Ludwig-Maximilians University, Munich, and at the South China Sea Institute of Oceanology. Magnetic remanence was measured on a 2G cryogenic magnetometer using the software of Wack (2010). Systematic thermal demagnetization steps were initially as large as 75 °C but decreased to 10 °C in and around the unblocking temperatures of pyrrhotite, magnetite, and/or hematite. Some sister samples were stepwise demagnetized using alternating field techniques up to peak alternating fields of 90 mT. PMGSC (R. Enkin) and PmagPy (Tauxe et al., 2016) software were used for principal component analyses (Kirschvink, 1980), great circle fitting (Halls, 1978), Fisher statistics (Fisher, 1953), and virtual geomagnetic pole calculations. Confidence limits were calculated according to Coe et al. (1985). We used the nonparametric fold test of Tauxe and Watson (1994) and the reversal test of McFadden and McElhinney (1990). To obtain a more accurate mean direction, we excluded the samples with maximum angular deviations exceeding 15° or when only great circles could be used to identify the magnetization component. Data are provided in Table S4.

3. Results

Thermal demagnetization of the volcanic rocks from the Lakang Formation isolates a high temperature component (HTC) that decays to the origin in the 280–350 and/or 400–580 °C interval (Figure 2a) after removing a magnetization component at lower unblocking temperatures that does not decay to the origin. Magnetization directions with single or both HTC unblocking intervals are similar (Figures 2a and 2b). Thermomagnetic (remanence and susceptibility) curves confirm the presence of magnetic phases with Curie temperatures of 325 and 580 °C (Figures S7 and S8) and, together with electron microprobe images and chemical analyses (Figure S3), demonstrate the presence of pyrrhotite and magnetite. Thermal and alternating field demagnetization trajectories yielded virtually identical directions (Figures 2b and 2c). Above 530 °C in some samples, magnetic susceptibility and/or remanent intensity begin to increase, and magnetization directions become erratic.
The HTC directions from the volcanic rocks are significantly different from the present-day field in geographic coordinates and display both polarities (Figure 1e). The reversal test is positive at the 95% confidence level (class C). The overall average of eight site mean directions is $D_g = 357.9°$, $I_g = 2.0°$ ($k_g = 5.8$, $\alpha_{95} = 25.1°$) and $D_s = 7.3°$, $I_s = −56.8°$ ($k_s = 12.9$, $\alpha_{95} = 16.1°$) ($D$, declination; $I$, inclination; $k$, Fisher precision parameter, $\alpha_{95}$, radius of a cone about the mean set at 95% confidence limits; $g$, geographic (in situ) coordinates; $s$, stratigraphic (tilt corrected) coordinates; Table S2). The fold test is positive with the optimal degree of untilting occurring at 110 ± 15°.

Demagnetization behavior of the Lakang limestones falls into three categories. (i) In 55% of the samples, a single component unblocks rapidly between 275 and 350 °C that decays toward the origin (Figure 2d). (ii)
In 35% of the samples, magnetic remanence drops sharply between 525 and 580 °C with magnetization directions that decay linearly to the origin (Figure 2e). (iii) In 10% of the samples, multiple magnetization components exist (Figure 2f). In category iii, the first magnetic component, which trends north and downward in geographic coordinates, is removed by 150–225 °C and does not trend toward the origin. Above 225 °C, remanent intensity decreases rapidly within two temperature intervals, 275–350 and 400–570 °C, with linear segments that most often decay toward the origin. The mean directions of the best-fit line segments from both temperature ranges are indistinguishable at 95% confidence limits (Table S2 and Figure S11), which together with thermomagnetic curves (Figures S7 and S8) and imaging and chemical analyses (Figure S4), indicate that both pyrrhotite and magnetite carry the remanence and that both magnetization components were locked-in at approximately the same time. For four sites that have both components, we used the magnetization directions unblocked in the 400–570 °C temperature range. The reversal test is positive at the 95% confidence level (class C). The fold test is positive, with a maximum at 103 ± 5% untilting. The average direction based on 20 site-mean directions is \( D_e = 352.2°, I_e = 1.3° (k_e = 3.2, \alpha_{95g} = 21.7°) \) and \( D_s = 6.4°, I_s = -57.8° (k_s = 30.8, \alpha_{95s} = 6.0°) \). An elongation/inclination (E/I) analysis (Tauxe & Kent, 2004) performed on 168 directions yields only a minor inclination correction (1.8°) with a mean inclination of \(-59.9°\) and best estimate between \(-68.2°\) and \(-57.3°\) at the 95% confidence level (Figure S12), indicating negligible inclination shallowing.

Because the tilt-corrected directions of the limestones and volcanic rocks are indistinguishable at 95% confidence limits (Figure S13), we combined both data sets to calculate the overall mean direction from the site-mean directions. Low temperature components are north and down in geographic coordinates, with an average direction of \( D_g = 0.3°, I_g = 46.6° (k_g = 24.7, \alpha_9 = 2.5°, n = 139) \)—quite similar to the present-day field (\( D = 359.8°, I = 43.6° \)) (Figure S14). Optimal clustering occurs at \(-3 ± 5%\) unfolding, signaling a modern-day-field viscous overprint. The overall HTC mean direction combining the 28 volcanic + carbonate sites is \( D_e = 354.0°, I_e = 1.5° (k_e = 3.8, \alpha_{95g} = 16.2°) \) and \( D_s = 6.7°, I_s = -57.5° (k_s = 23.2, \alpha_{95s} = 5.8°) \) (Figures 1e and 1f). The presence of dual magnetic polarity, positive field tests, and the roughly 13 Myr span in deposition (Zhu et al., 2008) suggests that the HTC represents sufficient time for adequate averaging of secular variation.

Stepwise thermal demagnetization of the sandstones and cherts from the western locality near Zhongba isolated two magnetic components in most samples (Figures 2g–2i). Below 225 °C, best fit directions not forced to the origin point north and down in geographic coordinates \( D_e = 358.4°, I_e = 49.5° (k_e = 36.7, \alpha_9 = 2.7°, n = 79) \), near the present-day field direction (\( D = 0.8°, I = 46.4° \)) (Figures 1e and 1f). The remanence of green-colored cherts decays toward the origin until a maximum temperature of 600 °C (Figure 2h), while the remanence of red-colored cherts decays toward the origin until 700 °C (Figure 2i). Rock magnetism and microscopy identify detrital iron oxides in both the sandstone and chert (Figures 5S and 5S7). Paleomagnetic directions from green and red cherts are indistinguishable, as are the directions from the sandstones and cherts (Figure S16), so we combined the data for further analyses.

The overall average HTC direction from 53 samples from the Rilang and Duobeng formations is \( D_e = 7.8°, I_e = -9.3° (k_e = 12.8, \alpha_9 = 5.7°) \) and \( D_s = 3.1°, I_s = -47.8° (k_s = 20.4, \alpha_9 = 4.4°) \) (Figures 1c and 1d), significantly different from the present-day field direction in geographic coordinates. The two stratigraphic sequences yield two normal polarity chron and one reverse chron within the Pianji section and one normal chron in the Duoping section (Tables S3 and S4). The reversal test is positive at the 95% confidence level (class C), while the nonparametric fold test gives optimal untilting at 105 ± 17% (\( k_s/k_g = 1.6 \)), yielding a positive fold test.

We combined the directions from both sections to apply the E/I correction method (Tauxe & Kent, 2004) to maximize the number of samples (\( n = 53 \)). The corrected inclination of \(-48.8°\) (95% confidence limits between \(-58.9°\) and \(-45.4°\), Figure S17) is within 1° of the non-E/I corrected inclination. Hence, inclination shallowing is insignificant, so we used the non-E/I-corrected direction with \( I_e = -47.8° \) in further analyses.
4. Discussion and Conclusions

The magnetizations of the rocks in both regions we studied yielded consistent, linear trajectories that decayed toward the origin with directions that pass reversal and fold tests. This is strong evidence favoring the hypothesis that the characteristic remanence was acquired at or very shortly after their formation. In addition, for the Lakang Formation, the characteristic directions display dispersion consistent with expected paleosecular variation from the TK03.GAD reference model (Tauxe & Kent, 2004) based on the S value of 21.7 and on the elongation parameter (E) of 1.5. The Lakang volcanic rocks display textures and mineral assemblages typical of oceanic basalts, where iron sulfides and iron oxides may coexist (Bach & Edwards, 2003). That both pyrrhotite and magnetite coexist in the Lakang limestones can be explained by primary formation of one of the phases that then partially or fully converted to the other. It is possible that heat and degassing/dewatering from the basalts altered the original magnetomineralogy of the limestones. The mean direction of the limestones coincides with that of the basalts, so if one phase was produced from alteration of the other, it must have happened close in time and was likely associated with the volcanism that was coeval with carbonate deposition. Because thermal demagnetization trajectories within the unblocking temperature range of both phases are the same, both magnetic phases formed faster than field reversal or plate motion time scales.

Debate surrounds the nature of magnetic remanence in Upper Cretaceous and Paleogene marine limestones from the Gamba region (Figure 1b), which also yielded positive fold and reversal tests with negligible inclination shallowing (Patzelt et al., 1996; Yi et al., 2011). A rock magnetic study on the same section led Huang et al. (2017) to conclude that the coexistence of pyrite and nanophase magnetite in the limestones signaled a secondary, chemical remagnetization. Yi et al. (2017) countered that the remanence was most likely acquired shortly after deposition or during early diagenesis and was therefore nearly equivalent to the deposition age of the rocks. Our results for the Lakang Formation reinforce this latter interpretation.

Sandstone and chert from the Zhongba region do not contain pyrrhotite and present a clear distinction between rock type and magnetic mineralogy. Remanence in the green cherts decays mostly within titanomagnetite Curie temperatures, while the remanence in red cherts unblocks within hematite Curie temperatures (Figures 2h and 2i). The magnetic remanence carriers in the sandstones are dominated by magnetite, and the remanence intensities of the sandstones are ~10–100 times higher than in the green cherts. These observations, together with the positive fold and reversal tests, lead us to conclude the magnetization in the Zhongba facies is primary.

We now consider the tectonic implications of our data assuming that they represent primary magnetizations, that they averaged secular variation and were not afflicted by inclination shallowing. The 137–125-Ma paleomagnetic poles for Zhongba and Cuona are 29.7°N, 260.1°E, A95 = 4.9° and 22.0°N, 266.7°E, A95 = 7.6°, with paleolatitudes of 30.5 ± 4.9°S and 39.6 ± 7.6°S, respectively. Southerly, temperate paleolatitudes are consistent with those proposed for Early Cretaceous bivalve fauna distributions from the Tethyan Himalaya (Rao et al., 2018) and formation of carbonate-rich sediments (Roberts et al., 2013). A tectonic reconstruction of Gondwanaland at 130 Ma (Torsvik et al., 2012) places India in the southern hemisphere rotated about 50° clockwise with respect to its present orientation (Figure 3c). Were Zhongba and Cuona rigidly attached to India in the Cretaceous, then this paleogeographic configuration would explain why the eastern locality of Cuona has a more southerly paleolatitude than the western locality of Zhongba. In other words, although they lie at similar paleolatitudes today, rotating India by ~50° clockwise would position the eastern locality of Cuona farther south of Zhongba. Such a conclusion appears independent of a comparison with Siberia—the western and eastern localities were 60.3 ± 4.5° and 65.9 ± 6.0° farther south with respect to Siberia at 130 Ma (Table S5).

However, India was positioned too far south to account for the Cuona and Zhongba paleolatitudes if they were at the same relative position with respect to India as seen today. Comparing the paleopoles from Zhongba and Cuona with the reference apparent polar wander path of the Indian plate (Torsvik et al., 2012) yields 16.6 ± 4.1° and 14.4 ± 5.9° of relative motion between them (Table S5), which should be the minimum estimate of N-S shortening within Greater India (Figure 3c). To resolve this discrepancy, we assume that Cuona and Zhongba were part of Greater India that subsequently became amalgamated to Asia sometime after the onset of the India-Asia collision as India subducted under Asia. Most of the relative motion between India and Asia was oriented ~N-S after 55 Ma (Figure 3b); hence, if Cuona and Zhongba
were never obducted onto Asia, they would lie somewhere north of their present locations under Asia. In other words, they would lie north of their present localities, somewhere roughly along longitude-par-allel great circle paths as shown in Figure 3a.

If we fix those same great circle paths to India in a 130-Ma reference frame, then the paleoposition of Cuona and Zhongba should lie at the intersection of their respective paleolatitudes and the great circles as shown in Figure 3c. As a result, Greater India would extend 1,950 ± 970 and 2,675 ± 720 km farther away from the present longitude of 92.4°E and 83.6°E, respectively (Figure 3). The uncertainties derived from the paleolatitude data that intersect the great circles yield the uncertainty for the extent of Greater India defined by these sites and are similar using apparent polar wander paths from Besse and Courtillot (2002) or Torsvik et al. (2012).

Our reconstruction matches well with paleomagnetic data derived from tectonic units that existed in eastern Gondwanaland during the Early Cretaceous. The Exmouth Plateau northwest of Australia contains Early Cretaceous sandstone and siltstone interbedded with mafic sills/flows that highly resembles the facies from Cuona (Gradstein & Ludden, 1992). Lower Cretaceous (125–136 Ma) rocks from the Gascoyne and Argo abyssal plains, near the Exmouth Plateau, yielded paleolatitudes of 36.6 ± 2.1°S and 36.7 ± 3.4°S (Ogg et al., 1992), respectively, that are quite compatible with Cuona. The Cape Range Fracture Zone, Wallaby-Zenith Fracture Zone, and Exmouth Plateau are all compatible with the paleogeography of Greater India shown in Figure 3c. No reconstruction other than Figure 3c can reconcile our paleomagnetic results and Early Cretaceous tectonic reconstructions of the southeastern Neo-Tethyan realm.

In the Tethyan Himalaya, Yang et al. (2015) and Ma et al. (2016) obtained paleolatitudes of 52.2° ± 5.7°S and 48.5 ± 6.1°S from 135- to 124-Ma basaltic lava flows in the Cuona and Langkazi regions. Positive fold and reversal tests indicated that the remanent magnetizations were primary. The significant differences in paleo-latitude of ours and Ma et al. (2016) and Yang et al. (2015) can be explained by differences in time when obduction occurred after India impinged against Asia in relation to where their site were located on the plate. Following the arguments of van Hinsbergen et al. (2012, 2018), our model is entirely consistent with the Ma et al. (2016) and Yang et al. (2015) results by assuming their localities were originally located much closer to the Indian craton and obducted much later (23 Ma, van Hinsbergen et al., 2018). On the other hand,

![Figure 3](image-url)
the elongation parameter (E) for both results are at odds with secular variation models that predict values of 1.3 for a latitude of 50°—being 3.1 in Yang et al. (2015) and 2.3 in Ma et al. (2016; Figure S18). Such highly elongated distributions suggest either a significant amount of flattening of the paleomagnetic directions (Tauxe & Kent, 2004) or, more likely for lava flows, that secular variation was inadequately averaged. Moreover, the presence of limestone at such high latitudes seems unlikely (Roberts et al., 2013), is incompatible with the paleomagnetic results from the Gascoyne and Argo abyssal plains (Ogg et al., 1992), and goes against Early Cretaceous bivalve fauna distributions from the Tethyan Himalaya (Rao et al., 2018).

Our new results define the minimum size of Greater India and argue that Greater India existed as a single entity with the rest of India since at least the Early Cretaceous. Greater India occupied an approximate surface area of ~4.7 × 10⁶ km², 2,000–2,700 km in the N-S direction and 2,500 km in the E-W direction, agreeing well with some estimates (Crawford, 1974; Ingalls et al., 2016; van Hinsbergen et al., 2018). Adopting the reconstruction for the southern margin of Asia following Meng et al. (2017), together with the reference curve for India (Torsvik et al., 2012) and accounting for the size of Greater India (Figure 3c), suggests that India first came into contact with Asia at ~55 Ma in the western part of the Himalayas (Figure 3b) and then proceeded diachronously across Tibet (Appel et al., 2012), although this latter interpretation is admittedly within uncertainty limits.

The onset of collision at ~55 Ma implies ~3,300-km convergence occurred between India and Asia. A >2,000-km extent of Greater India leads to intra-Asian shortening of ~1,000 km, consistent with recent paleomagnetic and geological estimates (Meng et al., 2017; van Hinsbergen et al., 2018), which is still only about one third of the N-S convergence since collision began, leaving large volumes of subducted material to facilitate plateau formation that doubled the thickness of the Tibetan crust in the Cenozoic (Capitanio et al., 2010; Molnar & Stock, 2009; Tapponnier et al., 2001). However, the composition (oceanic or continental) and thickness of Greater India’s lithosphere remains speculative. The Lagang Formation sandstone at Cuona represents the detritus of eroded continental material. Detrital zircons, petrology, and geochemistry of the Zhongba sandstone suggest affinities with Indian continental crust and correlate with other passive margin deposits in the Tethyan Himalaya, such as the sandstones in the Zansk Range of northern India, the Kumaon Himalayas, and the Thakkhola region of central Nepal (Du et al., 2015). Buoyant upper continental crust from the Greater Indian lithosphere could have been partially decoupled and stripped away from the down-going plate to become incorporated in the Himalayan fold-thrust belt, with the rest of the lithosphere being subducted (Ingalls et al., 2016). Thus, the large extent of Greater India, now defined quantitatively in our study, is compatible with models evoking crustal thickening via mass accumulation.

Our finding that Greater India existed since at least the Early Cretaceous and behaved as a single plate runs contrary to tectonic scenarios that split the oceanic basin at the leading edge of India into multiple plates, such as the Greater Indian Basin hypothesis of van Hinsbergen et al. (2012).

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