

1 **A record of plume-induced plate rotation triggering seafloor spreading and**
2 **subduction initiation**

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27 **The formation of a global network of plate boundaries surrounding a mosaic of**
28 **lithospheric fragments was a key step in the emergence of Earth's plate tectonics. So far,**
29 **propositions for plate boundary formation are regional in nature but how plate boundaries**
30 **are being created over 1000s of km in short periods of geological time remains elusive.**
31 **Here, we show from geological observations that a >12,000 km long plate boundary formed**
32 **between the Indian and African plates around 105 Ma with subduction segments from the**
33 **eastern Mediterranean region to a newly established India-Africa rotation pole in the west-**
34 **Indian ocean where it transitioned into a ridge between India and Madagascar. We find no**
35 **plate tectonics-related potential triggers of this plate rotation and identify coeval mantle**
36 **plume rise below Madagascar-India as the only viable driver. For this, we provide a proof**
37 **of concept by torque balance modeling revealing that the Indian and African cratonic keels**
38 **were important in determining plate rotation and subduction initiation in response to the**
39 **spreading plume head. Our results show that plumes may provide a non-plate-tectonic**
40 **mechanism for large plate rotation initiating divergent and convergent plate boundaries**
41 **far away from the plume head that may even be an underlying cause of the emergence of**
42 **modern plate tectonics.**

43 The early establishment of plate tectonics on Earth was likely a gradual process that
44 evolved as the cooling planet's lithosphere broke into a mosaic of major fragments, separated by
45 a network of plate boundaries: seafloor spreading ridges, transform faults, and subduction
46 zones¹. The formation of spreading ridges and connecting transform faults is regarded as a
47 passive process, occasionally associated with rising mantle plumes². The formation of
48 subduction zones is less well understood. Explanations for subduction initiation often infer
49 spontaneous gravitational collapse of aging oceanic lithosphere², or relocations of subduction
50 zones due to intraplate stress changes in response to continental collisions with other continents,
51 oceanic plateaus, or arcs³. Mantle plumes have also been suggested as drivers for regional
52 subduction initiation, primarily based on numerical modeling⁴⁻⁶. But while such processes may
53 explain how plate tectonics evolves on a regional scale, they do not provide insight into the
54 geodynamic cause(s) for the geologically sudden (<10 My) creation of often long (>1000 km)
55 plate boundaries including new subduction zones⁷. Demonstrating the causes of plate boundary
56 formation involving subduction initiation using the geological record is challenging and requires
57 (i) establishing whether subduction initiation was spontaneous or induced; (ii) if induced,

58 constraining the timing and direction of incipient plate convergence; (iii) reconstructing the
59 entire plate boundary from triple junction to triple junction, as well as the boundaries of
60 neighboring plates, to identify collisions, subduction terminations, or mantle plume arrival that
61 may have caused stress changes driving subduction initiation. In this paper, we provide such an
62 analysis for an intra-oceanic subduction zone that formed within the Neotethys ocean around 105
63 Ma, to evaluate the driver of subduction initiation and plate boundary formation.

64

65 **Induced subduction initiation across the Neotethys Ocean**

66 Determining spontaneous versus induced subduction initiation is a particular complexity
67 in this analysis and requires geological records of both the upper and lower plates: in both cases,
68 subduction initiation corresponds with initial lower plate burial, whereas coeval or delayed
69 extension in the upper plate are contrasting diagnostics of spontaneous or forced subduction
70 initiation, respectively⁸. Initiation of lower plate burial can be dated through prograde mineral
71 growth in rocks of the incipient subduction plate contact, in so-called metamorphic soles⁸. The
72 timing of extension is inferred from spreading records in so-called supra-subduction zone (SSZ)
73 ophiolites^{8-10,11}. Such SSZ ophiolites have a chemical stratigraphy widely interpreted as having
74 formed at spreading ridges above a nascent subduction zones. Metamorphic sole protoliths
75 typically reveal that also the initial downgoing plate was of oceanic composition^{2,9}, and so
76 ophiolite belts with metamorphic soles demarcate fossil juvenile intra-oceanic subduction plate
77 boundaries.

78 Several SSZ ophiolite belts exist in the Alpine-Himalayan mountain belt, which formed
79 during the closure of the Neotethys Ocean^{12,13} (Fig. 1A). One of these ophiolite belts formed in
80 Cretaceous time and runs from the eastern Mediterranean region to Pakistan, across northern
81 Arabia. The timing of lower plate burial as well as upper plate extension have been constrained
82 in this ophiolite belt through detailed geochronological, petrological, and geochemical work.
83 Incipient lower plate burial has been dated through Lu/Hf prograde garnet growth ages of ~104
84 Ma in Oman as well as in the eastern Mediterranean region^{8,14}. Upper plate extension and SSZ
85 ophiolite spreading has been dated using magmatic zircon U/Pb ages and synchronous
86 metamorphic sole ⁴⁰Ar/³⁹Ar cooling ages and occurred at 96-95 Ma (Pakistan, Oman)^{15,16} to 92-
87 90 Ma (Iran, eastern Mediterranean region)¹⁷. The 8-14 Myr time delay between initial lower

88 plate burial and upper plate extension demonstrates that initiation of this subduction zone was
89 not spontaneous, but induced by far-field forcing⁸.

90 An initial ~E-W convergence direction at this subduction zone was constrained through
91 paleomagnetic analysis and detailed kinematic reconstruction of post-subduction initiation
92 deformation of the eastern Mediterranean region, Oman, and Pakistan, and was accommodated at
93 ~N-S striking trench segments^{13,18-20}. This is surprising: for hundreds of Ma, throughout the
94 Tethyan realm rifts and ridges formed breaking fragments off northern Gondwana in the south,
95 which accreted at subduction zones to the southern Eurasian margin in the north^{21,22}. The ~E-W
96 convergence that triggered ~105 Ma subduction initiation across the Neotethys ocean was thus
97 near orthogonal to the long-standing plate motions. To find this trigger we developed the first
98 comprehensive reconstruction of the entire ~12,000 km long plate boundary that formed at ~105
99 Ma and placed this in context of reconstructions of collisions and mantle plumes of the
100 Neotethyan realm.

101

102 **Geological reconstruction of plate boundary formation across the Neotethys**

103 The Cretaceous SSZ ophiolites that formed at the Cretaceous intra-Neotethyan
104 subduction zone in its juvenile stages are now found as klippen on intensely deformed orogenic
105 belts (Fig. 1A). These belts formed during subduction zone migration and collisions with the
106 continents of Greater Adria, Arabia, and India. We reconstructed these orogenic belts (Fig. 1)
107 and restored the Cretaceous ophiolites into their original configuration (Fig. 1C) (see Methods).

108 The westernmost geological record of the Cretaceous intra-Neotethyan subduction zone
109 is found in eastern Greece and western Turkey, where it ended in a trench-trench-trench triple
110 junction with subduction zones along the southern Eurasian margin¹⁸. From there, east-dipping
111 (in the west) and west-dipping (in the east) subduction segments followed the saw-toothed shape
112 of the Greater Adriatic and Arabian continental margins (Fig. 1C) and initiated close to it: rocks
113 of these margins already underthrust the ophiolites within 5-15 My after SSZ ophiolite
114 spreading^{14,23,24}, and continent-derived zircons have been found in metamorphic sole rocks²⁵.
115 Subduction segments that likely nucleated along ancient N-S and NE-SW trending fracture
116 zones, linked through highly oblique, north-dipping subduction zones that trended parallel to and
117 likely reactivated the pre-existing (hyper)extended passive margins (Fig. 1B, C)^{20,23}. Subducted

118 remnants of the Cretaceous intra-Neotethyan subduction are well-resolved in the present-day
119 mantle as slabs below the southeastern Mediterranean Sea, central Arabia and the west Indian
120 Ocean²⁶.

121 East of Arabia, we trace the intra-oceanic plate boundary to a NE-SW striking, NW-
122 dipping subduction zone between the Kabul Block and the west Indian passive margin. The 96
123 Ma Waziristan ophiolites of Pakistan formed above this subduction zone and were thrust
124 eastward onto the Indian continental margin^{13,16} (Fig. 1B, C). This part of the plate boundary
125 may have inverted a spreading ridge that formed between the Kabul Block and India in the Early
126 Cretaceous¹³. The Cretaceous intra-Neotethyan plate boundary may have been convergent to as
127 far south as the Amirante Ridge in the west Indian Ocean¹³, but there is no record of
128 contemporaneous subduction beyond there. Instead, the plate boundary became extensional and
129 developed a rift, and later a mid-oceanic ridge in the Mascarene Basin that accommodated
130 separation of India from Madagascar^{13,27,28} (Fig. 1B). The plate boundary ended in a ridge-ridge-
131 ridge triple junction with ridges bordering the Antarctic plate in the south Indian Ocean^{13,28} (Fig.
132 1B).

133 The newly formed Cretaceous plate boundary essentially temporarily merged a large part
134 of Neotethyan oceanic lithosphere between Arabia and Eurasia to the Indian plate. This plate was
135 >12,000 km long from triple junction to triple junction, and reached from 45°S to 45°N, with
136 4500 km of rift/ridge in the southeast and 7500 km of subduction zone in the northwest and with
137 a transition between the convergent and divergent segments, representing the India-Africa Euler
138 pole¹³, in the west Indian Ocean (Fig. 1B). Marine geophysical constraints show a ~4°
139 counterclockwise rotation of India relative to Africa about the west Indian Ocean Euler pole
140 during rifting preceding the ~83 Ma onset of oceanic spreading in the Mascarene Basin²⁷⁻²⁹,
141 associated with up to hundreds of km of ~E-W convergence across the Neotethys (Fig. 1D).

142 The neighboring plates of the intra-Neotethyan subduction zone at 105 Ma were thus
143 Africa and India. The African plate was mostly surrounded by ridges and had a complex
144 subduction plate boundary in the Mediterranean region³⁰. The Indian plate was surrounded by
145 ridge-transform systems in the south and east and by subduction in the north, and may have
146 contained rifts and ridges between the Indian continent and Eurasia^{13,28}. The Neotethys
147 lithosphere between Arabia-Greater Adria and Eurasia continued unbroken to the north-dipping

148 subduction zone that had already existed along the southern Eurasian margin since the
149 Jurassic^{31,32}; the spreading ridges that existed during Neotethys Ocean opening in the Permian-
150 Triassic (north of Arabia)³³, and Triassic-Jurassic (eastern Mediterranean region)²³ had already
151 subducted below Eurasia by 105 Ma^{19,33} (Fig. 1B, C).

152

153 **Identifying potential drivers of plate boundary formation**

154 Collisions, subduction relocations, or mantle plume arrivals around or within the Indian
155 or African plates are all candidate processes to explain plate boundary formation at 105 Ma. At
156 the northern boundary of between these plates and southern Eurasia, many collisions of
157 microcontinents and arcs occurred since the Paleozoic, but none started or ended around 105
158 Ma^{13,21-23,33-35}. Continental subduction and collision was ongoing in the central Mediterranean
159 region²³, but it is not evident how this or any other changes in subduction dynamics along the E-
160 W trending southern Eurasian margin would lead to E-W convergence in the Neotethys Ocean.
161 In the eastern Neotethys, a mid-Cretaceous collision of the intra-oceanic Woyla Arc with the
162 Sundaland continental margin led to a subduction polarity reversal initiating eastward subduction
163 below Sundaland³⁶, which is recorded in ophiolites on the Andaman Islands. There, metamorphic
164 sole rocks with ⁴⁰Ar/³⁹Ar hornblende cooling ages of 105-106 Ma, and likely coeval SSZ
165 ophiolite spreading ages³⁷ reveal that this subduction zone may have developed slab pull around
166 the same time as the Indian Ocean-western Neotethys plate boundary formed (Fig 1C). However,
167 eastward slab pull below Sundaland cannot drive E-W convergence in the Neotethys to the west,
168 and Andaman SSZ extension may well be an expression rather than the trigger of Indian plate
169 rotation. Hence, we find no viable plate tectonics-related driver of the ~105 Ma plate boundary
170 formation that we reconstructed here.

171 A key role, however, is possible for the only remaining geodynamic, non-plate-tectonic,
172 plate-motion driver in the region: a mantle plume. India-Madagascar continental breakup is
173 widely viewed^{13,27,37} as related to the ~94 Ma and younger formation of the Morondava Large
174 Igneous Province (LIP) on Madagascar³⁸ and southwest India³⁹. This LIP, however, started
175 forming ~10 Ma after initial plate boundary formation. To understand whether the plume may be
176 responsible for both LIP emplacement and plate boundary formation, we conduct explorative
177 torque-balance simulations of plume-lithosphere interaction.

178

179 **Mantle plumes driving plate boundary formation and subduction initiation**

180 Numerical simulations of plume-lithosphere interaction have already identified that
181 plume head spreading below the lithosphere leads to horizontal asthenospheric flow that exerts a
182 ‘plume push’ force on the base of the lithosphere, particularly in the presence of a cratonic
183 keel^{5,40,41}. Plume push may accelerate plates by several cm/yr⁴¹ and has been proposed as a
184 potential driver of subduction initiation⁵.

185 In many cases, including in the case of the Morondava LIP, LIP eruption and
186 emplacement shortly preceded continental breakup, but pre-break up rifting preceded LIP
187 emplacement by 10-15 Myr²⁷. This early rifting typically is interpreted to indicate that the plume
188 migrated along the base of the lithosphere into a pre-existing rift that formed independently of
189 plume rise²⁷. However, in numerical simulations dynamic uplift⁴² and plume push⁴¹ already start
190 to accelerate plates 10-15 Myr before the plume head reaches the base of the lithosphere and
191 emplaces the LIP. Numerical simulations thus predict the observed delay between plume push,
192 as a driver for early rifting and subduction initiation, and LIP eruption and emplacement.

193 Here, we add to these plume-lithosphere coupling experiments by conducting proof-of-
194 concept torque-balance simulations particularly exploring why the observed India-Africa Euler
195 pole is so close to the plume head such that the associated plate rotation between Africa and
196 India caused E-W convergence in the Neotethys. We performed semi-analytical computations,
197 including both the Indian and African plates at ~105 Ma, and assess the influence of cratonic
198 keels on the position of the India-Africa Euler pole (Fig. 2, see Methods).

199 In our computations without cratonic keels, plume push under Madagascar/India caused
200 counterclockwise rotation of India versus Africa, but about an Euler pole situated far north of
201 Arabia, (Fig. 2A) without inducing significant E-W convergence within the Neotethys. However,
202 in experiments that include keels of the Indian and African cratonic lithosphere, which are
203 strongly coupled to the sub-asthenospheric mantle, the computed Euler pole location is shifted
204 southward towards the Indian continent, inducing E-W convergence along a larger part of the
205 plate boundary within the Neotethys Ocean (Fig. 2B).

206 Convergence of up to several hundreds of km, sufficient to induce self-sustaining
207 subduction²⁷, is obtained if plume material is fed into – and induced flow is confined to – a 200
208 km thick weak asthenospheric layer. The thinner this layer is, the further the plume head spreads,
209 and pushes the plate. The modern Indian cratonic root used in our computations has likely eroded
210 considerably during interaction with the ~70-65 Ma Deccan plume⁴³. India may have had a
211 thicker and/or laterally more extensive cratonic root at ~105 Ma than modeled here which would
212 further enhance coupling of the lithosphere and the sub-asthenospheric mantle. Furthermore, an
213 Euler pole close to India and a long convergent boundary to the north requires much weaker
214 coupling in the northern (oceanic) part of the India plate (Fig. 2). In this case, results remain
215 similar as long as the plume impinges near the southern part of the western boundary of
216 continental India.

217 An order of magnitude estimate of the maximum plume-induced stresses, assuming no
218 frictional resistance at other plate boundaries, is obtained from the rising force of $\sim 1.5 \cdot 10^{20}$ N of
219 a plume head with 1000 km diameter and density contrast 30 kg/m^3 . If half of this force acts on
220 the India plate and with a lever arm of 4000 km, this corresponds to a torque of $3 \cdot 10^{26}$ Nm. Once,
221 at the onset of rifting, ridge push is established as an additional force in the vicinity of the plume,
222 we estimate that this number may increase by up to a few tens of per cent. This torque can be
223 balanced at the convergent boundary (length ~ 5000 km, plate thickness ~ 100 km) involving
224 stresses of ~ 240 MPa, much larger than estimates of frictional resistance between subducting and
225 overriding plates that are only of the order of tens of MPa⁴⁴. For this estimate, we neglect any
226 frictional resistance at the base of the plate and at any other plate boundary – essentially
227 considering the plate as freely rotating above a pinning point. This is another endmember
228 scenario, as opposed to our above convergence estimate, where we had considered friction at the
229 plate base but neglected it at all plate boundaries. Therefore, the estimate of 240 MPa may be
230 considered as an upper bound but being compressive and oriented in the right direction it shows
231 the possibility of subduction initiation as has occurred in reality along the likely weakened
232 passive margin region of Arabia and Greater Adria. Moreover, the plume-induced compressive
233 stresses may have added to pre-existing compressive stresses, in particular due to ridge-push
234 around the African and Indian plates. Such additional compressive stresses may contribute to
235 shifting the Euler pole further south, closer to the position reconstructed in Fig. 1.

236 Subduction became self-sustained ~8-12 Ma after its initiation, as marked by the 96-92
237 Ma age of SSZ spreading^{15,17}: inception of this spreading shows that subduction rates exceeded
238 convergence rates, and reconstructed SSZ spreading rates were an order of magnitude higher¹⁵
239 than Africa-Arabia or Indian absolute plate motions^{41,45} signaling slab roll-back, i.e. self-
240 sustained subduction^{20,46}. Numerical models suggest that self-sustained subduction may start
241 after ~50-100 km of induced convergence⁷, corresponding to ~1° of India-Africa rotation
242 between ~105 and ~96-92 Ma. Subsequent east and west-dipping subduction segments (Fig. 1)
243 may have contributed to and accelerated the India-Africa/Arabia rotation, driving the
244 propagation of the Euler pole farther to the south (compare Fig. 2A, C).

245

246 **Mantle plumes as an initiator of plate tectonics?**

247 Previously, numerical modeling has shown that mantle plumes may trigger circular
248 subduction initiation around a plume head⁴, where local plume-related convection may drive
249 subduction of thermally weakened lithosphere. This subduction would propagate through slab
250 roll-back and may have started the first subduction features on Earth⁴. 3D convective models do
251 produce a global network of plate boundaries^{47,48} but the role of plumes in initiating new
252 subduction zones within this network is unclear. Here, we have provided the first evidence that
253 plume rise formed a >12,000 km long plate boundary composed of both convergent and
254 divergent segments. Our documented example is Cretaceous in age but geological observations
255 showing a general temporal overlap between LIP emplacement and formation of SSZ ophiolite
256 belts over more than a billion years⁴⁹ suggest that plume rise is a key driving factor in the
257 formation of subduction plate boundaries. Because mantle plumes are thought to be also
258 common features on planets without plate tectonics, such as Mars and Venus⁵⁰, they may have
259 played a vital role in the emergence of modern style plate tectonics on Earth. That plumes may
260 have been key for the evolution of plate tectonics on Earth, as we suggest, but apparently
261 insufficient on Mars and Venus, provides a new outlook on understanding the different planetary
262 evolutions.

263

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278 PJmcP, CGa, ELA and RLMV developed the kinematic reconstruction; BS performed
279 modelling; DJJvH, BS, CGu, WS wrote the paper, all authors made corrections and edits.

280

281 **Competing interests:** All authors declare no competing interests.

282

283

284 **Fig. 1.** Plate kinematic reconstructions of the Neotethys Ocean and surrounding continents at A)
285 the present-day; B) 70 Ma, corresponding to the time that most of the Neotethyan intra-oceanic
286 subduction zone had terminated due to arrival of the India, Africa-Arabia, and the Greater Adria
287 margin in the trench; C) 105 Ma, corresponding to the timing of intra-Neotethyan subduction
288 initiation and D) 110 Ma, just before intra-Neotethyan subduction initiation. An Euler pole
289 situated in the Indian Ocean north of Madagascar (yellow star) indicates the division between the
290 compressional plate boundary segment (the intra-Neotethys trench) and the extensional segment
291 (the incipient Mascarene rift connected to the mid-ocean ridge between Africa and Antarctica).
292 Rotation around this pole, and the related intra-Neotethyan subduction initiation, are interpreted
293 here to result from the rise and push of the Morondava mantle plume. See text for further
294 explanation, and Methods for the plate reconstruction approach and sources of detailed
295 restorations. Dark grey areas outline modern continents; light-grey areas indicate thinned
296 continental margins and microcontinents. Grey arrows indicate approximate rotational motion in
297 a mantle reference frame⁴⁵ around the Amirante Euler pole. AR = Amirante Ridge; Emed =
298 Eastern Mediterranean Region; Ir = Iran; LIP = Large Igneous Province; Mad = Madagascar;
299 Mas = Mascarene Basin; Pak = Pakistan, Tur = Turkey; Waz = Waziristan Ophiolite.

300

301 **Fig. 2.** The computed total displacement, induced by the Morondava plume (pink circle) for the
302 restored ~105 Ma plate configuration (Fig. 1C) for plates without (A, B) and with (C, D) African
303 and Indian cratonic keels, in an Africa-fixed (A, C), or mantle reference frame⁴⁵ (B, D) (see
304 Methods). It is assumed that, compared to a case with no lateral variations, the drag force due to
305 the plate moving over the mantle is increased by a factor of ten wherever reconstructed
306 lithosphere thickness exceeds 100 km (brown areas) and reduced to one tenth of the drag force
307 wherever it is less than 100 km thick. The India craton hence nearly “pins” the India plate, such
308 that its northern part moves in the opposite direction to the plume-induced push. Computation
309 assumes torque balance between plume push and shearing over asthenosphere; frictional
310 resistance at plate boundaries is neglected and computed convergence of several hundred km at
311 the northern end of the plate boundary is a maximum estimate. Ten degree grid spacing;

312 locations of plates, lithosphere thickness and the plume are reconstructed in a slab-fitted mantle
313 reference frame⁴⁵.

314

315 **Methods: Kinematic reconstruction** – The kinematic restoration of Neotethyan intra-
316 oceanic subduction was made in GPlates plate reconstruction software (www.gplates.org)⁵¹.
317 First, we systematically restored stable plates using marine geophysical data from the Atlantic
318 and Indian Ocean, and then restored continental margin deformation that occurred following the
319 arrival of continental lithosphere below the oceanic lithosphere preserved as ophiolites. These
320 restorations are based on a systematic reconstruction protocol, based on magnetic anomalies and
321 fracture zones of present-day sea floor and geophysical constraints on pre-drift extension in
322 adjacent passive continental margins²³, followed by kinematic restoration of post-obduction
323 orogenic deformation using structural geological constraints on continental extension, strike-slip
324 deformation, and shortening, and paleomagnetic constraints on vertical axis rotations. We then
325 restored pre-emplacement vertical axis microplate rotations^{52,53}, as well as paleo-orientations of
326 the SSZ spreading ridges at which the ophiolitic crust formed¹⁸⁻²⁰. The reconstruction shown in
327 Fig. 1B compiles kinematic restorations for the eastern Mediterranean region²³, Iran⁵⁴, Oman²⁰,
328 Pakistan¹³, and the Himalaya³⁴. Ophiolites interpreted to be part of the Cretaceous subduction
329 system include the 96-90 Ma, Cretaceous ophiolites exposed in SE Greece, Anatolia, Cyprus,
330 Syria, and Iraq, the Neyriz ophiolite of Iran, the Semail ophiolite in Oman, and the Waziristan-
331 Khost ophiolite in Pakistan and Afghanistan^{15-17,55}. The Jurassic ophiolite belts of northern
332 Turkey and Armenia⁵⁶⁻⁵⁸ and the late Cretaceous (<80 Ma) Kermanshah ophiolite of Iran⁵⁹ are
333 not included and are instead interpreted to have formed along the southern Eurasian margin²³.
334 The Masirah Ophiolite of East Oman⁶⁰ and the uppermost Cretaceous Bela, Muslim Bagh, and
335 Kabul-Altımur ophiolites of Pakistan and Afghanistan^{61,62} are interpreted to reflect oblique latest
336 Cretaceous to Paleogene India-Arabia convergence¹³ and are also unrelated to the event studied
337 here. Restoration of intra-oceanic subduction prior to the arrival of the continental margins used
338 paleomagnetic data from the ophiolites of Oman, Syria, Cyprus, and Turkey that constrain
339 vertical axis rotations, as well as the orientation of sheeted dyke following cooling after
340 intrusion^{18-20,52,53} as proxy for original ridge and intra-oceanic trench orientations. These
341 paleomagnetic data systematically revealed N-S to NW-SE primary sheeted dyke orientations¹⁸⁻

342 20,52,53. Because the ages of the SSZ ophiolites in the Neotethyan belt do not laterally progress,
343 spreading must have occurred near-orthogonal to the associated trench, which must thus also
344 have been striking N-S to NE-SW, as shown in the reconstruction of Fig. 1.

345 How far the Indian plate continued northwards around 105 Ma is subject to ongoing
346 debate. On the one hand, the northern Indian continental margin has been proposed to have rifted
347 off India sometime in the Cretaceous^{34,63}, but recent paleomagnetic data suggest that this process
348 occurred in the late Cretaceous, well after 100 Ma⁶⁴. Others inferred that the north Indian
349 continent had a passive margin contiguous with oceanic Neotethyan lithosphere since the middle
350 Jurassic or before and continued to a subduction zone below the SSZ ophiolites found in the
351 Himalayan suture zone and the Kohistan arc^{35,65,66}. Sedimentary and paleomagnetic data
352 demonstrate that these ophiolites formed adjacent to the Eurasian margin in the Early
353 Cretaceous⁶⁷, although they may have migrated southward during slab roll-back in the Late
354 Cretaceous³⁵. Recent paleomagnetic data have shown that a subduction zone may have existed
355 within the Neotethys to the west of the Andaman Islands, above which the West Burma Block
356 would have been located (Figure 1)⁶⁸. Our reconstruction of the eastern Neotethys may thus be
357 oversimplified. However, the geological record of the West Burma Block shows that this
358 subduction zone already existed as early as 130 Ma, and E-W trending until well into the
359 Cenozoic⁶⁸, and we see no reason to infer that changes in the eastern Neotethys contributed to
360 the plate boundary formation discussed here. Some have speculated that the West Burma
361 subduction zone would have been connected to a long-lived, equatorial subduction zone within
362 the Neotethys all along the Indian segment that would already have existed in the Early
363 Cretaceous⁶⁹: this scenario remains unconstrained by paleomagnetic data, and is inconsistent
364 with sediment provenance data from the Himalaya and overlying ophiolites³⁵. In summary, the
365 Indian plate around 105 Ma continued far into the Neotethyan realm, and the India-Africa
366 rotation is a likely driver of E-W convergence sparking subduction initiation close to the
367 northern Gondwana margin purported in Figure 1.

368 *Torque balance modeling* – Forces considered here include (i) the push due to plume-
369 induced flow in the asthenosphere and (ii) the drag due to shear flow between the moving plate
370 and a deeper mantle at rest (Fig. S1). In the first case, we disregard any lateral variations. Plume-
371 induced flow is treated as Poiseuille flow, i.e. with parabolic flow profile, in an asthenospheric
372 channel of thickness h_c , radially away from the plume stem. Since at greater distance plume-

373 induced flow will eventually not remain confined to the asthenosphere, we only consider it to a
 374 distance 2400 km, in accord with numerical results⁴¹, and consistent with the finding that there is
 375 a transition from dominantly pressure-driven Poiseuille flow at shorter wavelengths to
 376 dominantly shear-driven Couette flow at length scales approximately exceeding mantle
 377 depth^{70,71}. With v_0 the velocity in the center of the channel at a distance d from the plume stem
 378 the total volume flux rate is $2/3 \cdot v_0 \cdot 2\pi d \cdot h_c$ (here neglecting the curvature of the Earth surface
 379 for simplicity). Its time integral is equal to the volume of the plume head with radius estimated⁷²
 380 to be about $r_p=500$ km, with considerable uncertainty. That is, integration is done over a time
 381 interval until the entire plume head volume has flown into the asthenospheric channel. Hence the
 382 corresponding displacement vector in the center of the channel is

$$\mathbf{x}_{plu} = \int_{\Delta t} v_0 dt \cdot \mathbf{e}_r = \frac{r_p^3}{d \cdot h_c} \cdot \mathbf{e}_r$$

383
 384 where \mathbf{e}_r is the unit vector radially away from the plume (red arrows in Extended Data Fig. 1).
 385 Because of the parabolic flow profile, the vertical displacement gradient at the top of the channel
 386 is

$$2 \cdot \frac{\mathbf{x}_{plu}}{0.5 \cdot h_c} = 2 \cdot \int_{\Delta t} v_0 dt \cdot \frac{1}{0.5 \cdot h_c} \cdot \mathbf{e}_r = \frac{4r_p^3}{d \cdot h_c^2} \cdot \mathbf{e}_r.$$

387
 388 Viscosity is defined such that the force per area is equal to viscosity times the radial gradient of
 389 horizontal velocity. Hence the time integral of torque on the plate is

$$\mathbf{T}_{plu} = \frac{4\eta_0}{h_c} \int_A \mathbf{r} \times \mathbf{x}_{plu} dA = \frac{4\eta_0 r_p^3}{d \cdot h_c^2} \int_A \mathbf{r} \times \mathbf{e}_r dA$$

390
 391 where η_0 is viscosity in the channel and \mathbf{r} is the position vector. \mathbf{T}_{plu} is balanced by the time-
 392 integrated torque \mathbf{T}_{pla} of the plate rotating an angle ω over the underlying mantle. With plate
 393 displacement vectors $\mathbf{x}_{pla} = \omega \times \mathbf{r}$ (black arrows in Fig. S1) we obtain

$$\mathbf{T}_{pla} = -\frac{\eta_0}{h_s} \int_A \mathbf{r} \times \mathbf{x}_{pla} dA = -\frac{\eta_0}{h_s} \int_A \mathbf{r} \times (\omega \times \mathbf{r}) dA$$

394

395 Here h_s is an effective thickness of the layer over which shearing occurs, which is calculated
 396 below for a stratified viscosity structure, i.e. laterally homogeneous coupling of plate and mantle
 397 and which we will set equal to h_c for simplicity. Specifically, with \mathbf{T}_x being the time-integrated
 398 torque acting on a plate rotating an angle ω_0 around the x-axis

$$\mathbf{T}_x = -\frac{\omega_0 \eta_0}{h_s} \int_A \mathbf{r} \times (\mathbf{e}_x \times \mathbf{r}) dA,$$

399

400 and \mathbf{T}_y and \mathbf{T}_z defined in analogy, the torque balance equation can be written

$$\mathbf{T}_{plu} = \frac{\omega_x}{\omega_0} \cdot \mathbf{T}_x + \frac{\omega_y}{\omega_0} \cdot \mathbf{T}_y + \frac{\omega_z}{\omega_0} \cdot \mathbf{T}_z$$

401

402 ω_0 cancels out when \mathbf{T}_x , \mathbf{T}_y and \mathbf{T}_z are inserted. Integrals used to compute these torques only
 403 depend on plate geometry, η_0 cancels out in the torque balance, and we can solve for the rotation
 404 angle vector $\boldsymbol{\omega}$ simply by a 3 x 3 matrix inversion. In the more general case, where we do not set
 405 h_s and h_c equal, $\boldsymbol{\omega}$ is scaled by a factor h_s/h_c .

406 If a plate moves over a mantle where viscosity varies with depth, then the force per area
 407 F/A should be the same at all depths, and the radial gradient of horizontal velocity $dv/dz = F/A \cdot$
 408 $1/\eta(z)$. If we assume that the deep mantle is at rest (i.e. it moves slowly compared to plate
 409 motions), we further find that plate motion is

$$v_0 = \int_{z_0}^{z(\eta_{\max})} \frac{dv}{dz} dz = \frac{F}{A} \int_{z_0}^{z(\eta_{\max})} \frac{1}{\eta(z)} dz =: \frac{F h_s}{A \eta_0} \quad (1)$$

410

411 The integration is done from the base of the lithosphere z_0 to the depth where the approximation
 412 of the “mantle at rest” is probably the most closely matched, i.e. we choose the viscosity
 413 maximum. The last equality is according to the definition of the effective layer thickness,
 414 whereby η_0 is the viscosity just below the lithosphere. Solving this equation for h_s for the
 415 viscosity structure in Extended Data Fig. 2 and a 100 km thick lithosphere gives $h_s=203.37$ km.

416 The plume location at 27.1°E, 40.4° S, is obtained by rotating the center of the
 417 corresponding LIP at 46° E, 26° S and an age 87 Ma (adopted from Doubrovine et al.⁷³) in the

418 slab-fitted mantle reference frame⁴⁵, in which also the plate geometries at 105 Ma are
419 reconstructed.

420 Results for this case (Fig. 2A) show that a plume pushing one part of a plate may induce
421 a rotation of that plate, such that other parts of that plate may move in the opposite direction. A
422 simple analog is a sheet of paper pushed, near its bottom left corner, to the right: Then, near the
423 top left corner, the sheet will move to the left. With two sheets (plates) on either side, local
424 divergence near the bottom (near the plume) may turn into convergence near the top (at the part
425 of the plate boundary furthest away from the plume). The length of that part of the plate
426 boundary, where convergence is induced may increase, if one plate is nearly “pinned” at a hinge
427 point slightly NE of the plume, perhaps due to much stronger coupling between plate and mantle.
428 At the times considered here ~105 My ago, the Indian continent, where coupling was presumably
429 stronger, was in the southern part of the Indian plate, whereas in its north, there was a large
430 oceanic part, with presumably weaker coupling. Hence the geometry was indeed such that
431 convergence could be induced along a longer part of the plate boundary.

432 In the second case, we therefore consider lateral variations in the coupling between plate
433 and mantle, corresponding to variations in lithosphere thickness and/or asthenosphere viscosity,
434 by multiplying the drag force (from the first case) at each location with a resistance factor. This
435 factor is a function of lithosphere thickness reconstructed at 105 Ma. On continents, thickness
436 derived from tomography⁷⁴ with slabs removed⁷⁵ is simply backward-rotated. In the oceans, we
437 use thickness [km] = $10 \cdot (\text{age [Ma]} - 105)^{0.5}$ with ages from present-day Earthbyte age grid
438 version 3.6, i.e. accounting for the younger age and reduced thickness at 105 Ma, besides
439 backward-rotating. To determine the appropriate rotation, the lithosphere (in present-day
440 location) is divided up into India, Africa, Arabia, Somalia and Madagascar (paleo-)plates and
441 respective 105 Ma finite rotations from van der Meer et al.⁴⁵ are applied. For the parts of the
442 reconstructed plates where thickness could not be reconstructed in this way – often, because this
443 part of the plate has been subducted – we first extrapolate thickness up to a distance $\sim 2.3^\circ$, and
444 set the thickness to a default value of 80 km for the remaining part. Reconstructed thickness is
445 shown in Extended Data Fig. 4. For the resistance factor as a function of lithosphere thickness
446 we use two models: Firstly, we use a continuous curve (Extended Data Fig. 3) according to eq.
447 (1)

$$\frac{F}{A} = \frac{v_0}{\int_{z_0}^{z(\eta_{\max})} \frac{1}{\eta(z)} dz}. \quad (2)$$

448

449 with the mantle viscosity model in Extended Data Fig. 2 combined with variable lithosphere
450 thickness z_0 . However, this causes only a minor change in the plate rotations (Extended Data Fig.
451 4 compared to Fig. 2B). Hence, we also use a stronger variation, further explained in the caption
452 of Fig 2 and with results shown in Fig. 2C and D.

453

454 **Data availability**

455 GPlates files with reconstructions used to draft Figure 1 are provided at
456 https://figshare.com/articles/dataset/van_Hinsbergen_NatureGeo_2021_GPlates_zip/13516727.

457

458 **Code availability**

459 All codes used in the geodynamic modeling in this study are available at
460 https://figshare.com/articles/software/van_Hinsbergen_etal_NatureGeo_2021_geodynamics_package/13635089.

462

463

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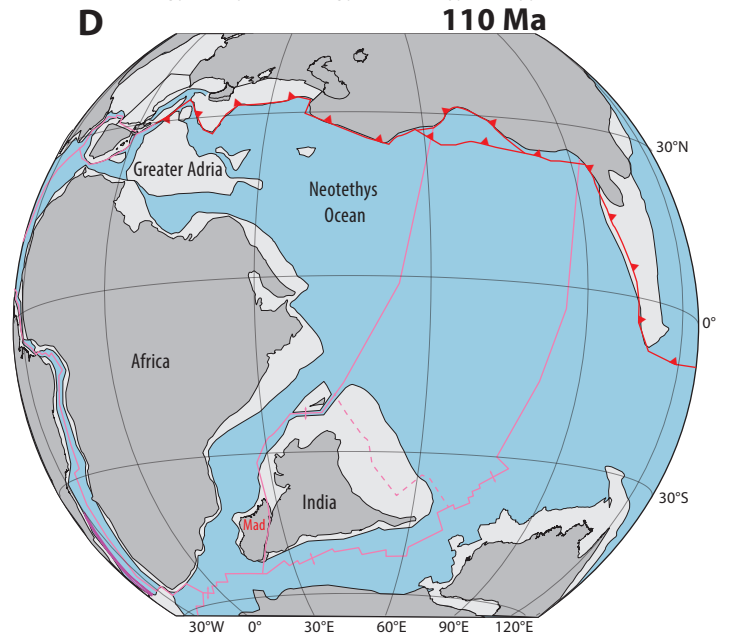
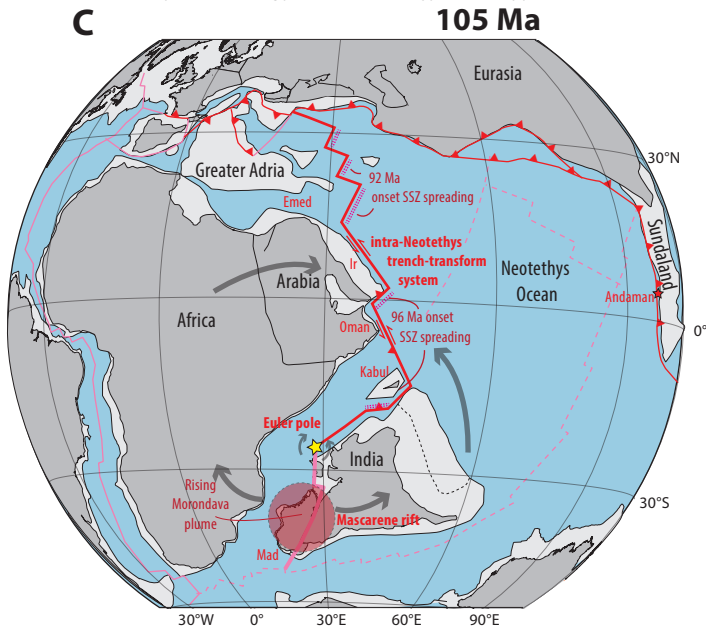
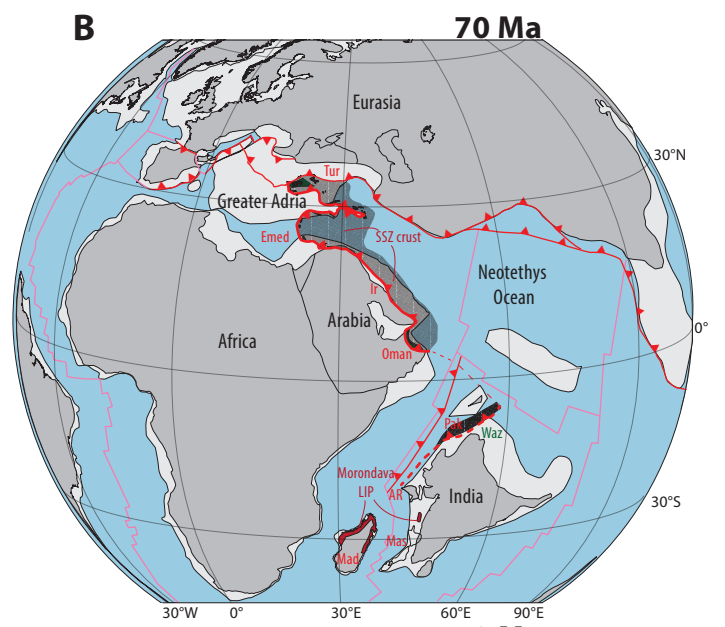
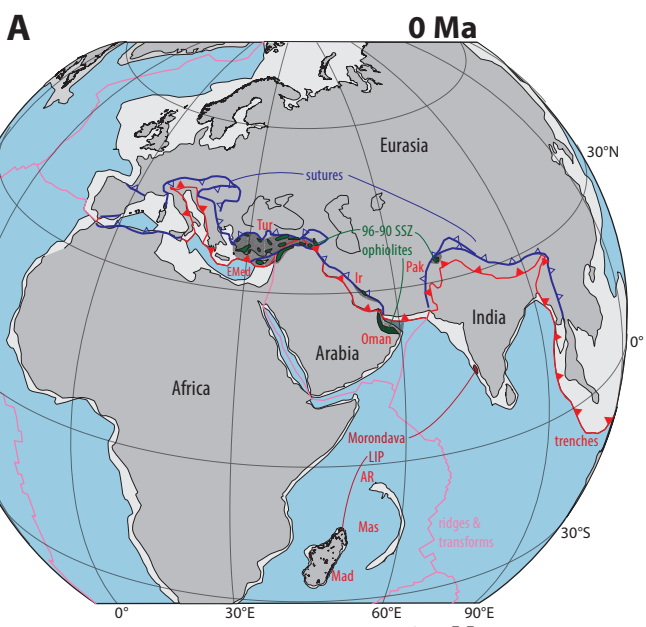
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