

## A record of plume-induced plate rotation triggering subduction initiation

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DOI:

[10.1038/s41561-021-00780-7](https://doi.org/10.1038/s41561-021-00780-7)

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*Document Version*

Peer reviewed version

*Citation for published version (Harvard):*

van Hinsbergen, D, Steinberger, B, Guilmette, C, Maffione, M, Gürer, D, Peters, K, Plunder, A, McPhee, PJ, Gaina, C, Advokaat, E, Vissers, RLM & Spakman, W 2021, 'A record of plume-induced plate rotation triggering subduction initiation', *Nature Geoscience*, vol. 14, no. 8, pp. 626-630. <https://doi.org/10.1038/s41561-021-00780-7>

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van Hinsbergen, D.J.J., Steinberger, B., Guilmette, C. et al. A record of plume-induced plate rotation triggering subduction initiation. *Nat. Geosci.* (2021). <https://doi.org/10.1038/s41561-021-00780-7>

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

3  
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23 **The formation of a global network of plate boundaries surrounding a mosaic of**  
24 **lithospheric fragments was a key step in the emergence of Earth's plate tectonics. So far,**  
25 **propositions for plate boundary formation are regional in nature but how plate boundaries**  
26 **are being created over 1000s of km in short periods of geological time remains elusive.**  
27 **Here, we show from geological observations that a >12,000 km long plate boundary formed**  
28 **between the Indian and African plates around 105 Ma with subduction segments from the**  
29 **eastern Mediterranean region to a newly established India-Africa rotation pole in the west-**  
30 **Indian ocean where it transitioned into a ridge between India and Madagascar. We find no**  
31 **plate tectonics-related potential triggers of this plate rotation and identify coeval mantle**  
32 **plume rise below Madagascar-India as the only viable driver. For this, we provide a proof**  
33 **of concept by torque balance modeling revealing that the Indian and African cratonic keels**  
34 **were important in determining plate rotation and subduction initiation in response to the**  
35 **spreading plume head. Our results show that plumes may provide a non-plate-tectonic**  
36 **mechanism for large plate rotation initiating divergent and convergent plate boundaries**  
37 **far away from the plume head that may even be an underlying cause of the emergence of**  
38 **modern plate tectonics.**

39 The early establishment of plate tectonics on Earth was likely a gradual process that  
40 evolved as the cooling planet's lithosphere broke into a mosaic of major fragments, separated by  
41 a network of plate boundaries: seafloor spreading ridges, transform faults, and subduction  
42 zones<sup>1</sup>. The formation of spreading ridges and connecting transform faults is regarded as a  
43 passive process, occasionally associated with rising mantle plumes<sup>2</sup>. The formation of  
44 subduction zones is less well understood. Explanations for subduction initiation often infer  
45 spontaneous gravitational collapse of aging oceanic lithosphere<sup>2</sup>, or relocations of subduction  
46 zones due to intraplate stress changes in response to continental collisions with other continents,  
47 oceanic plateaus, or arcs<sup>3</sup>. Mantle plumes have also been suggested as drivers for regional  
48 subduction initiation, primarily based on numerical modeling<sup>4-6</sup>. But while such processes may  
49 explain how plate tectonics evolves on a regional scale, they do not provide insight into the  
50 geodynamic cause(s) for the geologically sudden (<10 My) creation of often long (>1000 km)  
51 plate boundaries including new subduction zones<sup>7</sup>. Demonstrating the causes of plate boundary  
52 formation involving subduction initiation using the geological record is challenging and requires  
53 (i) establishing whether subduction initiation was spontaneous or induced; (ii) if induced,

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

60

### 61 **Induced subduction initiation across the Neotethys Ocean**

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

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

84 plate burial and upper plate extension demonstrates that initiation of this subduction zone was  
85 not spontaneous, but induced by far-field forcing<sup>8</sup>.

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

97

### 98 **Geological reconstruction of plate boundary formation across the Neotethys**

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

104 The westernmost geological record of the Cretaceous intra-Neotethyan subduction zone  
105 is found in eastern Greece and western Turkey, where it ended in a trench-trench-trench triple  
106 junction with subduction zones along the southern Eurasian margin<sup>18</sup>. From there, east-dipping  
107 (in the west) and west-dipping (in the east) subduction segments followed the saw-toothed shape  
108 of the Greater Adriatic and Arabian continental margins (Fig. 1C) and initiated close to it: rocks  
109 of these margins already underthrust the ophiolites within 5-15 My after SSZ ophiolite  
110 spreading<sup>14,23,24</sup>, and continent-derived zircons have been found in metamorphic sole rocks<sup>25</sup>.  
111 Subduction segments that likely nucleated along ancient N-S and NE-SW trending fracture  
112 zones, linked through highly oblique, north-dipping subduction zones that trended parallel to and  
113 likely reactivated the pre-existing (hyper)extended passive margins (Fig. 1B, C)<sup>20,23</sup>. Subducted

114 remnants of the Cretaceous intra-Neotethyan subduction are well-resolved in the present-day  
115 mantle as slabs below the southeastern Mediterranean Sea, central Arabia and the west Indian  
116 Ocean<sup>26</sup>.

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

129 The newly formed Cretaceous plate boundary essentially temporarily merged a large part  
130 of Neotethyan oceanic lithosphere between Arabia and Eurasia to the Indian plate. This plate was  
131 >12,000 km long from triple junction to triple junction, and reached from 45°S to 45°N, with  
132 4500 km of rift/ridge in the southeast and 7500 km of subduction zone in the northwest and with  
133 a transition between the convergent and divergent segments, representing the India-Africa Euler  
134 pole<sup>13</sup>, in the west Indian Ocean (Fig. 1B). Marine geophysical constraints show a ~4°  
135 counterclockwise rotation of India relative to Africa about the west Indian Ocean Euler pole  
136 during rifting preceding the ~83 Ma onset of oceanic spreading in the Mascarene Basin<sup>27,29</sup>,  
137 associated with up to hundreds of km of ~E-W convergence across the Neotethys (Fig. 1D).

138 The neighboring plates of the intra-Neotethyan subduction zone at 105 Ma were thus  
139 Africa and India. The African plate was mostly surrounded by ridges and had a complex  
140 subduction plate boundary in the Mediterranean region<sup>30</sup>. The Indian plate was surrounded by  
141 ridge-transform systems in the south and east and by subduction in the north, and may have  
142 contained rifts and ridges between the Indian continent and Eurasia<sup>13,28</sup>. The Neotethys  
143 lithosphere between Arabia-Greater Adria and Eurasia continued unbroken to the north-dipping

144 subduction zone that had already existed along the southern Eurasian margin since the  
145 Jurassic<sup>31,32</sup>; the spreading ridges that existed during Neotethys Ocean opening in the Permian-  
146 Triassic (north of Arabia)<sup>33</sup>, and Triassic-Jurassic (eastern Mediterranean region)<sup>23</sup> had already  
147 subducted below Eurasia by 105 Ma<sup>19,33</sup> (Fig. 1B, C).

148

### 149 **Identifying potential drivers of plate boundary formation**

150 Collisions, subduction relocations, or mantle plume arrivals around or within the Indian  
151 or African plates are all candidate processes to explain plate boundary formation at 105 Ma. At  
152 the northern boundary of between these plates and southern Eurasia, many collisions of  
153 microcontinents and arcs occurred since the Paleozoic, but none started or ended around 105  
154 Ma<sup>13,21-23,33-35</sup>. Continental subduction and collision was ongoing in the central Mediterranean  
155 region<sup>23</sup>, but it is not evident how this or any other changes in subduction dynamics along the E-  
156 W trending southern Eurasian margin would lead to E-W convergence in the Neotethys Ocean.  
157 In the eastern Neotethys, a mid-Cretaceous collision of the intra-oceanic Woyla Arc with the  
158 Sundaland continental margin led to a subduction polarity reversal initiating eastward subduction  
159 below Sundaland<sup>36</sup>, which is recorded in ophiolites on the Andaman Islands. There, metamorphic  
160 sole rocks with <sup>40</sup>Ar/<sup>39</sup>Ar hornblende cooling ages of 105-106 Ma, and likely coeval SSZ  
161 ophiolite spreading ages<sup>37</sup> reveal that this subduction zone may have developed slab pull around  
162 the same time as the Indian Ocean-western Neotethys plate boundary formed (Fig 1C). However,  
163 eastward slab pull below Sundaland cannot drive E-W convergence in the Neotethys to the west,  
164 and Andaman SSZ extension may well be an expression rather than the trigger of Indian plate  
165 rotation. Hence, we find no viable plate tectonics-related driver of the ~105 Ma plate boundary  
166 formation that we reconstructed here.

167 A key role, however, is possible for the only remaining geodynamic, non-plate-tectonic,  
168 plate-motion driver in the region: a mantle plume. India-Madagascar continental breakup is  
169 widely viewed<sup>13,27,37</sup> as related to the ~94 Ma and younger formation of the Morondava Large  
170 Igneous Province (LIP) on Madagascar<sup>38</sup> and southwest India<sup>39</sup>. This LIP, however, started  
171 forming ~10 Ma after initial plate boundary formation. To understand whether the plume may be  
172 responsible for both LIP emplacement and plate boundary formation, we conduct explorative  
173 torque-balance simulations of plume-lithosphere interaction.

174

## 175 **Mantle plumes driving plate boundary formation and subduction initiation**

176 Numerical simulations of plume-lithosphere interaction have already identified that  
177 plume head spreading below the lithosphere leads to horizontal asthenospheric flow that exerts a  
178 ‘plume push’ force on the base of the lithosphere, particularly in the presence of a cratonic  
179 keel<sup>5,40,41</sup>. Plume push may accelerate plates by several cm/yr<sup>41</sup> and has been proposed as a  
180 potential driver of subduction initiation<sup>5</sup>.

181 In many cases, including in the case of the Morondava LIP, LIP eruption and  
182 emplacement shortly preceded continental breakup, but pre-break up rifting preceded LIP  
183 emplacement by 10-15 Myr<sup>27</sup>. This early rifting typically is interpreted to indicate that the plume  
184 migrated along the base of the lithosphere into a pre-existing rift that formed independently of  
185 plume rise<sup>27</sup>. However, in numerical simulations dynamic uplift<sup>42</sup> and plume push<sup>41</sup> already start  
186 to accelerate plates 10-15 Myr before the plume head reaches the base of the lithosphere and  
187 emplaces the LIP. Numerical simulations thus predict the observed delay between plume push,  
188 as a driver for early rifting and subduction initiation, and LIP eruption and emplacement.

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

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



202 Convergence of up to several hundreds of km, sufficient to induce self-sustaining  
203 subduction<sup>27</sup>, is obtained if plume material is fed into – and induced flow is confined to – a 200  
204 km thick weak asthenospheric layer. The thinner this layer is, the further the plume head spreads,  
205 and pushes the plate. The modern Indian cratonic root used in our computations has likely eroded  
206 considerably during interaction with the ~70-65 Ma Deccan plume<sup>43</sup>. India may have had a  
207 thicker and/or laterally more extensive cratonic root at ~105 Ma than modeled here which would  
208 further enhance coupling of the lithosphere and the sub-asthenospheric mantle. Furthermore, an  
209 Euler pole close to India and a long convergent boundary to the north requires much weaker  
210 coupling in the northern (oceanic) part of the India plate (Fig. 2). In this case, results remain  
211 similar as long as the plume impinges near the southern part of the western boundary of  
212 continental India.

213 An order of magnitude estimate of the maximum plume-induced stresses, assuming no  
214 frictional resistance at other plate boundaries, is obtained from the rising force of  $\sim 1.5 \cdot 10^{20}$  N of  
215 a plume head with 1000 km diameter and density contrast  $30 \text{ kg/m}^3$ . If half of this force acts on  
216 the India plate and with a lever arm of 4000 km, this corresponds to a torque of  $3 \cdot 10^{26}$  Nm. Once,  
217 at the onset of rifting, ridge push is established as an additional force in the vicinity of the plume,  
218 we estimate that this number may increase by up to a few tens of per cent. This torque can be  
219 balanced at the convergent boundary (length ~5000 km, plate thickness ~100 km) involving  
220 stresses of ~240 MPa, much larger than estimates of frictional resistance between subducting and  
221 overriding plates that are only of the order of tens of MPa<sup>44</sup>. For this estimate, we neglect any  
222 frictional resistance at the base of the plate and at any other plate boundary – essentially  
223 considering the plate as freely rotating above a pinning point. This is another endmember  
224 scenario, as opposed to our above convergence estimate, where we had considered friction at the  
225 plate base but neglected it at all plate boundaries. Therefore, the estimate of 240 MPa may be  
226 considered as an upper bound but being compressive and oriented in the right direction it shows  
227 the possibility of subduction initiation as has occurred in reality along the likely weakened  
228 passive margin region of Arabia and Greater Adria. Moreover, the plume-induced compressive  
229 stresses may have added to pre-existing compressive stresses, in particular due to ridge-push  
230 around the African and Indian plates. Such additional compressive stresses may contribute to  
231 shifting the Euler pole further south, closer to the position reconstructed in Fig. 1.

232 Subduction became self-sustained ~8-12 Ma after its initiation, as marked by the 96-92  
233 Ma age of SSZ spreading<sup>15,17</sup>: inception of this spreading shows that subduction rates exceeded  
234 convergence rates, and reconstructed SSZ spreading rates were an order of magnitude higher<sup>15</sup>  
235 than Africa-Arabia or Indian absolute plate motions<sup>41,45</sup> signaling slab roll-back, i.e. self-  
236 sustained subduction<sup>20,46</sup>. Numerical models suggest that self-sustained subduction may start  
237 after ~50-100 km of induced convergence<sup>7</sup>, corresponding to ~1° of India-Africa rotation  
238 between ~105 and ~96-92 Ma. Subsequent east and west-dipping subduction segments (Fig. 1)  
239 may have contributed to and accelerated the India-Africa/Arabia rotation, driving the  
240 propagation of the Euler pole farther to the south (compare Fig. 2A, C).

241

### 242 **Mantle plumes as an initiator of plate tectonics?**

243 Previously, numerical modeling has shown that mantle plumes may trigger circular  
244 subduction initiation around a plume head<sup>4</sup>, where local plume-related convection may drive  
245 subduction of thermally weakened lithosphere. This subduction would propagate through slab  
246 roll-back and may have started the first subduction features on Earth<sup>4</sup>. 3D convective models do  
247 produce a global network of plate boundaries<sup>47,48</sup> but the role of plumes in initiating new  
248 subduction zones within this network is unclear. Here, we have provided the first evidence that  
249 plume rise formed a >12,000 km long plate boundary composed of both convergent and  
250 divergent segments. Our documented example is Cretaceous in age but geological observations  
251 showing a general temporal overlap between LIP emplacement and formation of SSZ ophiolite  
252 belts over more than a billion years<sup>49</sup> suggest that plume rise is a key driving factor in the  
253 formation of subduction plate boundaries. Because mantle plumes are thought to be also  
254 common features on planets without plate tectonics, such as Mars and Venus<sup>50</sup>, they may have  
255 played a vital role in the emergence of modern style plate tectonics on Earth. That plumes may  
256 have been key for the evolution of plate tectonics on Earth, as we suggest, but apparently  
257 insufficient on Mars and Venus, provides a new outlook on understanding the different planetary  
258 evolutions.

259

260 **Acknowledgments:** DJJvH, MM, DG, AP, and ELA were funded through European Research  
261 Council Starting Grant 306810 (SINK) to DJJvH. DJJvH, KP and PJMcP were funded

262 through Netherlands Organization for Scientific Research (NWO) Vidi grant 864.11.004  
263 to DJJvH. DJJvH acknowledges Netherlands Organization for Scientific Research  
264 (NWO) Vici grant 865.17.001. BS and CGa received funding from the Research Council  
265 of Norway through its Centres of Excellence funding scheme, project number 223272.  
266 BS received additional funding from the innovation pool of the Helmholtz Association  
267 through the “Advanced Earth System Modelling Capacity (ESM)” activity. CG was  
268 funded through Discovery Grant (RGPIN-2014-05681) from the National Science and  
269 Engineering Research Council of Canada. We thank Inge Loes ten Kate and Debaditya  
270 Bandyopadhyay for discussion, and Fabio Capitanio, Dietmar Müller, and an anonymous  
271 reviewer for their constructive comments.

272

273 **Author contributions:** DJJvH, BS, WS designed research. DJJvH, CGu, MM, DG, KP, AP,  
274 PJmcP, CGa, ELA and RLMV developed the kinematic reconstruction; BS performed  
275 modelling; DJJvH, BS, CGu, WS wrote the paper, all authors made corrections and edits.

276

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

278

279

280 **Fig. 1.** Plate kinematic reconstructions of the Neotethys Ocean and surrounding continents at A)  
281 the present-day; B) 70 Ma, corresponding to the time that most of the Neotethyan intra-oceanic  
282 subduction zone had terminated due to arrival of the India, Africa-Arabia, and the Greater Adria  
283 margin in the trench; C) 105 Ma, corresponding to the timing of intra-Neotethyan subduction  
284 initiation and D) 110 Ma, just before intra-Neotethyan subduction initiation. An Euler pole  
285 situated in the Indian Ocean north of Madagascar (yellow star) indicates the division between the  
286 compressional plate boundary segment (the intra-Neotethys trench) and the extensional segment  
287 (the incipient Mascarene rift connected to the mid-ocean ridge between Africa and Antarctica).  
288 Rotation around this pole, and the related intra-Neotethyan subduction initiation, are interpreted  
289 here to result from the rise and push of the Morondava mantle plume. See text for further  
290 explanation, and Methods for the plate reconstruction approach and sources of detailed  
291 restorations. Dark grey areas outline modern continents; light-grey areas indicate thinned  
292 continental margins and microcontinents. Grey arrows indicate approximate rotational motion in  
293 a mantle reference frame<sup>45</sup> around the Amirante Euler pole. AR = Amirante Ridge; Emed =  
294 Eastern Mediterranean Region; Ir = Iran; LIP = Large Igneous Province; Mad = Madagascar;  
295 Mas = Mascarene Basin; Pak = Pakistan, Tur = Turkey; Waz = Waziristan Ophiolite.

297 **Fig. 2.** The computed total displacement, induced by the Morondava plume (pink circle) for the  
298 restored ~105 Ma plate configuration (Fig. 1C) for plates without (A, B) and with (C, D) African  
299 and Indian cratonic keels, in an Africa-fixed (A, C), or mantle reference frame<sup>45</sup> (B, D) (see  
300 Methods). It is assumed that, compared to a case with no lateral variations, the drag force due to  
301 the plate moving over the mantle is increased by a factor of ten wherever reconstructed  
302 lithosphere thickness exceeds 100 km (brown areas) and reduced to one tenth of the drag force  
303 wherever it is less than 100 km thick. The India craton hence nearly “pins” the India plate, such  
304 that its northern part moves in the opposite direction to the plume-induced push. Computation  
305 assumes torque balance between plume push and shearing over asthenosphere; frictional  
306 resistance at plate boundaries is neglected and computed convergence of several hundred km at  
307 the northern end of the plate boundary is a maximum estimate. Ten degree grid spacing;  
308 locations of plates, lithosphere thickness and the plume are reconstructed in a slab-fitted mantle  
309 reference frame<sup>45</sup>.

310

311 **Methods: Kinematic reconstruction** – The kinematic restoration of Neotethyan intra-  
312 oceanic subduction was made in GPlates plate reconstruction software ([www.gplates.org](http://www.gplates.org))<sup>51</sup>.  
313 First, we systematically restored stable plates using marine geophysical data from the Atlantic  
314 and Indian Ocean, and then restored continental margin deformation that occurred following the  
315 arrival of continental lithosphere below the oceanic lithosphere preserved as ophiolites. These  
316 restorations are based on a systematic reconstruction protocol, based on magnetic anomalies and  
317 fracture zones of present-day sea floor and geophysical constraints on pre-drift extension in  
318 adjacent passive continental margins<sup>23</sup>, followed by kinematic restoration of post-obduction  
319 orogenic deformation using structural geological constraints on continental extension, strike-slip  
320 deformation, and shortening, and paleomagnetic constraints on vertical axis rotations. We then  
321 restored pre-emplacement vertical axis microplate rotations<sup>52,53</sup>, as well as paleo-orientations of  
322 the SSZ spreading ridges at which the ophiolitic crust formed<sup>18-20</sup>. The reconstruction shown in  
323 Fig. 1B compiles kinematic restorations for the eastern Mediterranean region<sup>23</sup>, Iran<sup>54</sup>, Oman<sup>20</sup>,  
324 Pakistan<sup>13</sup>, and the Himalaya<sup>34</sup>. Ophiolites interpreted to be part of the Cretaceous subduction  
325 system include the 96-90 Ma, Cretaceous ophiolites exposed in SE Greece, Anatolia, Cyprus,

326 Syria, and Iraq, the Neyriz ophiolite of Iran, the Semail ophiolite in Oman, and the Waziristan-  
327 Khost ophiolite in Pakistan and Afghanistan<sup>15-17,55</sup>. The Jurassic ophiolite belts of northern  
328 Turkey and Armenia<sup>56-58</sup> and the late Cretaceous (<80 Ma) Kermanshah ophiolite of Iran<sup>59</sup> are  
329 not included and are instead interpreted to have formed along the southern Eurasian margin<sup>23</sup>.  
330 The Masirah Ophiolite of East Oman<sup>60</sup> and the uppermost Cretaceous Bela, Muslim Bagh, and  
331 Kabul-Altimur ophiolites of Pakistan and Afghanistan<sup>61,62</sup> are interpreted to reflect oblique latest  
332 Cretaceous to Paleogene India-Arabia convergence<sup>13</sup> and are also unrelated to the event studied  
333 here. Restoration of intra-oceanic subduction prior to the arrival of the continental margins used  
334 paleomagnetic data from the ophiolites of Oman, Syria, Cyprus, and Turkey that constrain  
335 vertical axis rotations, as well as the orientation of sheeted dyke following cooling after  
336 intrusion<sup>18-20,52,53</sup> as proxy for original ridge and intra-oceanic trench orientations. These  
337 paleomagnetic data systematically revealed N-S to NW-SE primary sheeted dyke orientations<sup>18-  
338 20,52,53</sup>. Because the ages of the SSZ ophiolites in the Neotethyan belt do not laterally progress,  
339 spreading must have occurred near-orthogonal to the associated trench, which must thus also  
340 have been striking N-S to NE-SW, as shown in the reconstruction of Fig. 1.

341 How far the Indian plate continued northwards around 105 Ma is subject to ongoing  
342 debate. On the one hand, the northern Indian continental margin has been proposed to have rifted  
343 off India sometime in the Cretaceous<sup>34,63</sup>, but recent paleomagnetic data suggest that this process  
344 occurred in the late Cretaceous, well after 100 Ma<sup>64</sup>. Others inferred that the north Indian  
345 continent had a passive margin contiguous with oceanic Neotethyan lithosphere since the middle  
346 Jurassic or before and continued to a subduction zone below the SSZ ophiolites found in the  
347 Himalayan suture zone and the Kohistan arc<sup>35,65,66</sup>. Sedimentary and paleomagnetic data  
348 demonstrate that these ophiolites formed adjacent to the Eurasian margin in the Early  
349 Cretaceous<sup>67</sup>, although they may have migrated southward during slab roll-back in the Late  
350 Cretaceous<sup>35</sup>. Recent paleomagnetic data have shown that a subduction zone may have existed  
351 within the Neotethys to the west of the Andaman Islands, above which the West Burma Block  
352 would have been located (Figure 1)<sup>68</sup>. Our reconstruction of the eastern Neotethys may thus be  
353 oversimplified. However, the geological record of the West Burma Block shows that this  
354 subduction zone already existed as early as 130 Ma, and E-W trending until well into the  
355 Cenozoic<sup>68</sup>, and we see no reason to infer that changes in the eastern Neotethys contributed to  
356 the plate boundary formation discussed here. Some have speculated that the West Burma

357 subduction zone would have been connected to a long-lived, equatorial subduction zone within  
 358 the Neotethys all along the Indian segment that would already have existed in the Early  
 359 Cretaceous<sup>69</sup>: this scenario remains unconstrained by paleomagnetic data, and is inconsistent  
 360 with sediment provenance data from the Himalaya and overlying ophiolites<sup>35</sup>. In summary, the  
 361 Indian plate around 105 Ma continued far into the Neotethyan realm, and the India-Africa  
 362 rotation is a likely driver of E-W convergence sparking subduction initiation close to the  
 363 northern Gondwana margin purported in Figure 1.

364 *Torque balance modeling* – Forces considered here include (i) the push due to plume-  
 365 induced flow in the asthenosphere and (ii) the drag due to shear flow between the moving plate  
 366 and a deeper mantle at rest (Fig. S1). In the first case, we disregard any lateral variations. Plume-  
 367 induced flow is treated as Poiseuille flow, i.e. with parabolic flow profile, in an asthenospheric  
 368 channel of thickness  $h_c$ , radially away from the plume stem. Since at greater distance plume-  
 369 induced flow will eventually not remain confined to the asthenosphere, we only consider it to a  
 370 distance 2400 km, in accord with numerical results<sup>41</sup>, and consistent with the finding that there is  
 371 a transition from dominantly pressure-driven Poiseuille flow at shorter wavelengths to  
 372 dominantly shear-driven Couette flow at length scales approximately exceeding mantle  
 373 depth<sup>70,71</sup>. With  $v_0$  the velocity in the center of the channel at a distance  $d$  from the plume stem  
 374 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  
 375 for simplicity). Its time integral is equal to the volume of the plume head with radius estimated<sup>72</sup>  
 376 to be about  $r_p=500$  km, with considerable uncertainty. That is, integration is done over a time  
 377 interval until the entire plume head volume has flown into the asthenospheric channel. Hence the  
 378 corresponding displacement vector in the center of the channel is

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

379  
 380 where  $\mathbf{e}_r$  is the unit vector radially away from the plume (red arrows in Extended Data Fig. 1).  
 381 Because of the parabolic flow profile, the vertical displacement gradient at the top of the channel  
 382 is

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

383

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

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

386  
 387 where  $\eta_0$  is viscosity in the channel and  $\mathbf{r}$  is the position vector.  $\mathbf{T}_{plu}$  is balanced by the time-  
 388 integrated torque  $\mathbf{T}_{pla}$  of the plate rotating an angle  $\boldsymbol{\omega}$  over the underlying mantle. With plate  
 389 displacement vectors  $\mathbf{x}_{pla} = \boldsymbol{\omega} \times \mathbf{r}$  (black arrows in Fig. S1) we obtain

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

390  
 391 Here  $h_s$  is an effective thickness of the layer over which shearing occurs, which is calculated  
 392 below for a stratified viscosity structure, i.e. laterally homogeneous coupling of plate and mantle  
 393 and which we will set equal to  $h_c$  for simplicity. Specifically, with  $\mathbf{T}_x$  being the time-integrated  
 394 torque acting on a plate rotating an angle  $\omega_0$  around the x-axis

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

395  
 396 and  $\mathbf{T}_y$  and  $\mathbf{T}_z$  defined in analogy, the torque balance equation can be written

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

397  
 398  $\boldsymbol{\omega}_0$  cancels out when  $\mathbf{T}_x$ ,  $\mathbf{T}_y$  and  $\mathbf{T}_z$  are inserted. Integrals used to compute these torques only  
 399 depend on plate geometry,  $\eta_0$  cancels out in the torque balance, and we can solve for the rotation  
 400 angle vector  $\boldsymbol{\omega}$  simply by a 3 x 3 matrix inversion. In the more general case, where we do not set  
 401  $h_s$  and  $h_c$  equal,  $\boldsymbol{\omega}$  is scaled by a factor  $h_s/h_c$ .

402 If a plate moves over a mantle where viscosity varies with depth, then the force per area  
 403  $F/A$  should be the same at all depths, and the radial gradient of horizontal velocity  $dv/dz = F/A \cdot$   
 404  $1/\eta(z)$ . If we assume that the deep mantle is at rest (i.e. it moves slowly compared to plate  
 405 motions), we further find that plate motion is



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

406

407 The integration is done from the base of the lithosphere  $z_0$  to the depth where the approximation  
 408 of the “mantle at rest” is probably the most closely matched, i.e. we choose the viscosity  
 409 maximum. The last equality is according to the definition of the effective layer thickness,  
 410 whereby  $\eta_0$  is the viscosity just below the lithosphere. Solving this equation for  $h_s$  for the  
 411 viscosity structure in Extended Data Fig. 2 and a 100 km thick lithosphere gives  $h_s=203.37$  km.

412 The plume location at 27.1°E, 40.4° S, is obtained by rotating the center of the  
 413 corresponding LIP at 46° E, 26° S and an age 87 Ma (adopted from Doubrovine et al.<sup>73</sup>) in the  
 414 slab-fitted mantle reference frame<sup>45</sup>, in which also the plate geometries at 105 Ma are  
 415 reconstructed.

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

428 In the second case, we therefore consider lateral variations in the coupling between plate  
 429 and mantle, corresponding to variations in lithosphere thickness and/or asthenosphere viscosity,  
 430 by multiplying the drag force (from the first case) at each location with a resistance factor. This  
 431 factor is a function of lithosphere thickness reconstructed at 105 Ma. On continents, thickness  
 432 derived from tomography<sup>74</sup> with slabs removed<sup>75</sup> is simply backward-rotated. In the oceans, we  
 433 use thickness [km] = 10 · (age [Ma] - 105)<sup>0.5</sup> with ages from present-day Earthbyte age grid

434 version 3.6, i.e. accounting for the younger age and reduced thickness at 105 Ma, besides  
435 backward-rotating. To determine the appropriate rotation, the lithosphere (in present-day  
436 location) is divided up into India, Africa, Arabia, Somalia and Madagascar (paleo-)plates and  
437 respective 105 Ma finite rotations from van der Meer et al.<sup>45</sup> are applied. For the parts of the  
438 reconstructed plates where thickness could not be reconstructed in this way – often, because this  
439 part of the plate has been subducted – we first extrapolate thickness up to a distance  $\sim 2.3^\circ$ , and  
440 set the thickness to a default value of 80 km for the remaining part. Reconstructed thickness is  
441 shown in Extended Data Fig. 4. For the resistance factor as a function of lithosphere thickness  
442 we use two models: Firstly, we use a continuous curve (Extended Data Fig. 3) according to eq.  
443 (1)

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

444  
445 with the mantle viscosity model in Extended Data Fig. 2 combined with variable lithosphere  
446 thickness  $z_0$ . However, this causes only a minor change in the plate rotations (Extended Data Fig.  
447 4 compared to Fig. 2B). Hence, we also use a stronger variation, further explained in the caption  
448 of Fig 2 and with results shown in Fig. 2C and D.

449

#### 450 **Data availability**

451 GPlates files with reconstructions used to draft Figure 1 are provided at  
452 [https://figshare.com/articles/dataset/van\\_Hinsbergen\\_NatureGeo\\_2021\\_GPlates\\_zip/13516727](https://figshare.com/articles/dataset/van_Hinsbergen_NatureGeo_2021_GPlates_zip/13516727).

453

#### 454 **Code availability**

455 All codes used in the geodynamic modeling in this study are available at  
456 [https://figshare.com/articles/software/van\\_Hinsbergen\\_etal\\_NatureGeo\\_2021\\_geodynamics\\_package/13635089](https://figshare.com/articles/software/van_Hinsbergen_etal_NatureGeo_2021_geodynamics_package/13635089).

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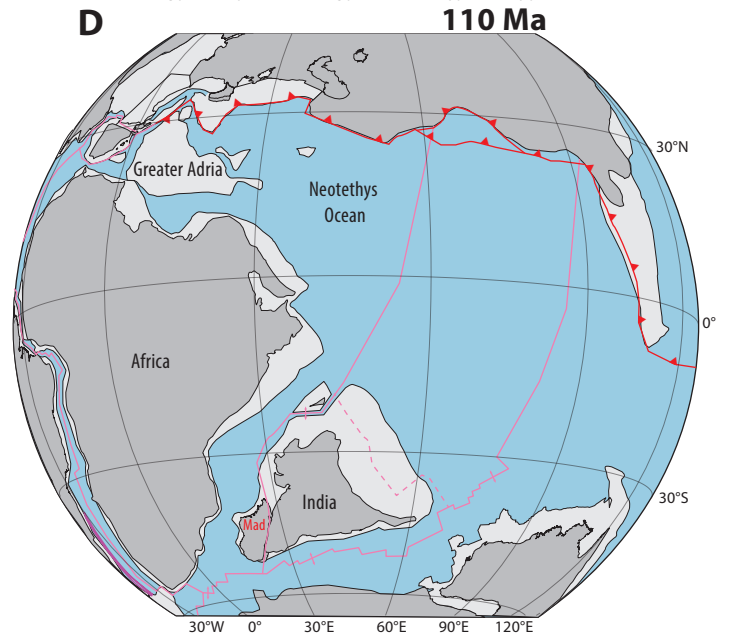
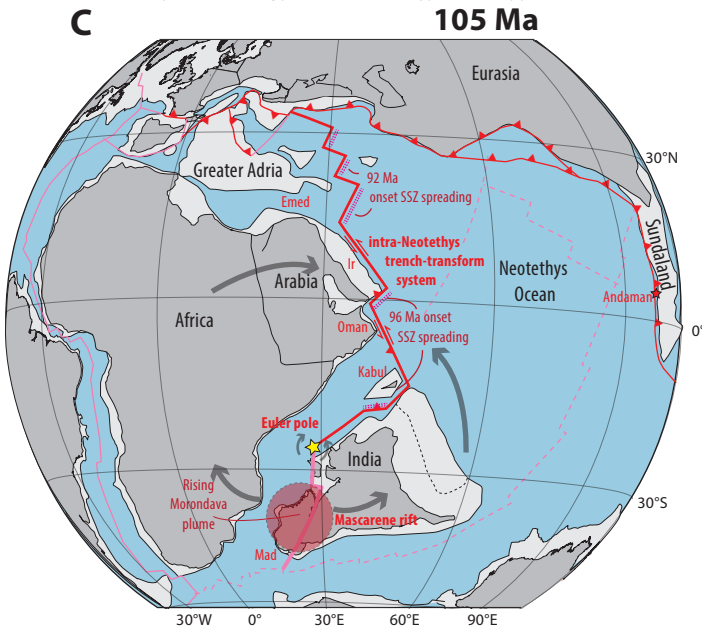
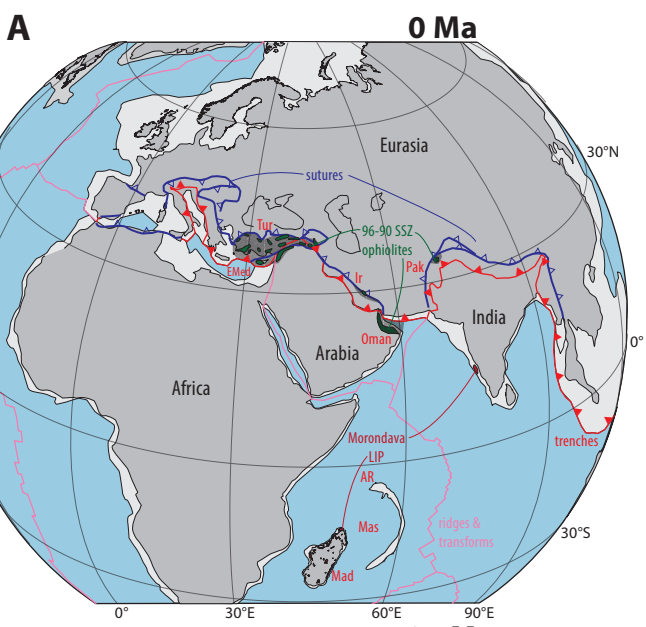
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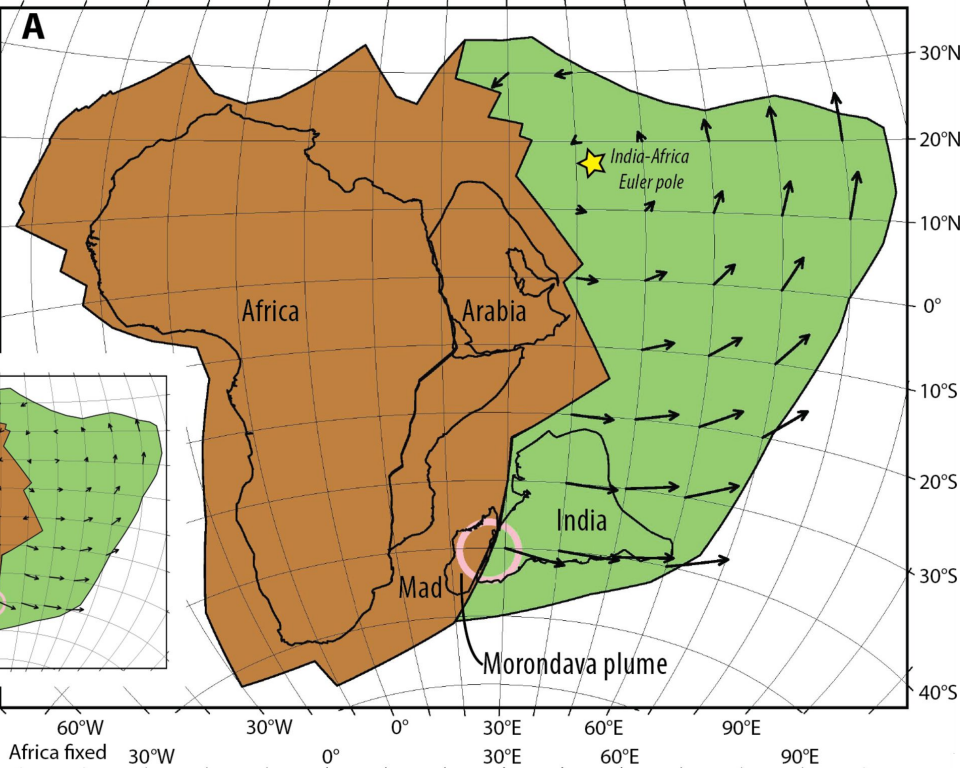
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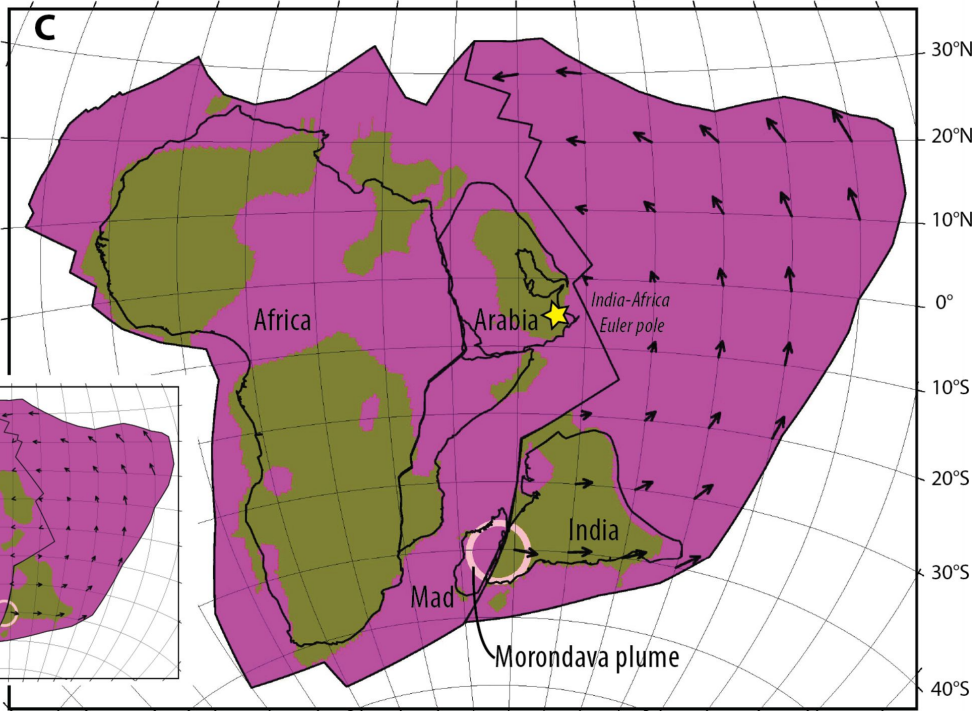
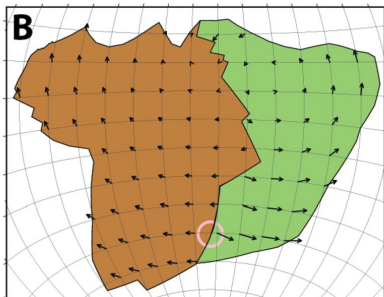
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Africa fixed



Mantle frame



Mantle frame

