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1 **Neogene South Asian Monsoon Rainfall and Wind Histories diverged due to**
2 **topography effects**

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23 **The drivers of South Asian Monsoon evolution remain highly debated. An**
24 **intensification of monsoonal rainfall recorded in terrestrial and marine sediment**
25 **archives from the earliest Miocene (23-20 million years ago, Ma) is generally**
26 **attributed to Himalayan uplift. However, Indian Ocean paleorecords place the onset**
27 **of a strong monsoon around 13 Ma, linked to strengthening of the southwesterly**
28 **winds of the Somali Jet that also force Arabian Sea upwelling. Here we reconcile**
29 **these divergent records using Earth System Model simulations to evaluate the**
30 **interactions between paleogeography and ocean-atmosphere dynamics. We show**
31 **that factors forcing South Asian Monsoon circulation versus rainfall are decoupled**
32 **and diachronous. Himalayan and Tibetan Plateau topography predominantly**
33 **controlled early Miocene rainfall patterns, with limited impact on ocean-**

34 **atmosphere circulation. Yet the uplift of East African and Middle Eastern**
35 **topography played a pivotal role in the establishment of modern Somali Jet**
36 **structure above the western Indian Ocean, while strong upwelling initiates as a**
37 **direct consequence of the Arabian Peninsula emergence and the initiation of**
38 **modern-like atmospheric circulation. Our results emphasize that although elevated**
39 **rainfall seasonality was likely a persistent feature since the India-Asia collision in**
40 **the Paleogene, the modern-like monsoonal atmospheric circulation was only**
41 **reached in the late Neogene.**

42 The South Asian monsoon (SAM) is a key element of Asian climate that sustains
43 populations over a vast region *via* continental rainfall and wind-driven coastal upwelling
44 supporting important marine ecosystem services. Assessing past SAM evolution and
45 underlying mechanisms is therefore essential to understand its behavior in warm climate
46 conditions and to better constrain feedbacks with topography and global climate on
47 geological timescales¹. A sound understanding of factors controlling past monsoonal
48 changes is, however, hampered by apparently conflicting interpretations of monsoon
49 proxies²⁻¹⁰. Isotopic and botanical records indicate the existence of strong seasonal SAM
50 rainfall on land since at least the Paleogene^{3,4}, while sedimentary archives of monsoon-
51 driven erosion and weathering in the Himalayan foreland and northern Indian Ocean
52 suggest an important intensification of monsoonal rainfall around 25-20 million years ago
53 (Ma), peaking at 15 Ma⁵. In contrast, a late middle Miocene (~13 Ma) onset of the
54 'modern-like' SAM has been inferred from ocean sediment archives in the western
55 Arabian Sea (ODP sites 722 and 730)⁶⁻¹⁰ and Maldives archipelago (IODP expedition

56 359)¹¹ (Fig. 1a). Those records indicate the onset of strong wind-driven coastal upwelling
57 in the western Arabian Sea, associated with high primary production and oxygen
58 minimum zone, and a reorganization of tropical Indian Ocean surface circulation
59 attributed to the inception of low-level winds with a strength and/or pattern similar to today
60 (Fig. 1a). The mechanism(s) driving this relatively late appearance of ‘modern-like’
61 monsoon circulation^{6,7,11}, well after Himalayan uplift^{12,13} that is nonetheless generally
62 accepted as the main driver of SAM intensification^{5,6,13-14}, remains uncertain.

63 The apparent decoupling between monsoon rainfall and winds during the early to middle
64 Miocene (23-11 Ma) either suggests that regional proxy records are not all faithfully
65 recording monsoon strength or that we need to reassess the widely-held view that
66 Himalaya-Tibetan Plateau elevation is the primary control on both SAM wind and rainfall
67 intensity^{17,18}. Previous modeling studies highlighted the sensitivity of the SAM to regional
68 topography^{17,19-22} and advanced our mechanistic understanding of SAM forcing and the
69 complex relationship between atmospheric circulation, rainfall, and orography¹⁷. Yet the
70 degree of coupling between these parameters in the past remains poorly understood
71 because, to date, such studies have not considered land-sea distribution despite its
72 important role in controlling monsoon dynamics²³⁻²⁴. Additionally, past atmospheric
73 dynamics, surface ocean circulation and marine biogeochemistry need to be assessed
74 together to enable realistic comparisons with marine records used to trace past monsoon
75 circulation.

76 Here we investigate both SAM evolution and biological productivity in the western Arabian
77 sea during the Miocene with an Earth System Model (IPSL-CM5A2²⁵), combined with an
78 offline ocean biogeochemistry component (PISCES-v2²⁶) (see Methods). We performed
79 the experiments using realistic land-sea configurations at ~20 Ma (early Miocene,
80 hereafter EM) and ~10 Ma (late Miocene, hereafter LM) (See Methods, Fig. 1 and
81 Extended Data Figure 1). This allows us to assess the relationship between the Somali
82 Jet, the dominant low-level wind pattern above the Arabian Sea (Fig. 1a), and western
83 Arabian Sea upwelling during the Miocene. We then investigate the potential influence of
84 Miocene paleogeographic and global climate changes on Somali Jet structure and SAM
85 rainfall intensity using sensitivity experiments. Our results reveal that drivers of ocean-
86 atmosphere dynamical changes in the western Arabian sea are decoupled from drivers
87 of rainfall patterns in continental Asia since the Miocene.

88 **Miocene upwelling and primary productivity**

89 Our simulations show that primary productivity in the western Arabian Sea increases (Fig.
90 2a) and becomes strongly seasonal (Fig. 2b) in response to paleogeographic changes
91 from the early to late Miocene. In the EM simulation, primary productivity in the surface
92 ocean (0-40 meters depth) is limited by low nutrient availability (Fig. 2b), restricting
93 biological production to the deep photic zone (60-120m) where sufficient nutrients and
94 light are available (Fig. 2c). In contrast, primary productivity in the LM simulation is highly
95 seasonal, with maximum values during late boreal summer (August-September) (Fig. 2b)
96 sustained by increased nutrient delivery to the euphotic zone relative to EM (Fig. 2b-c).

97 The seasonal availability of nutrients therefore imposes a substantial constraint on
98 summer productivity in this region and its enhancement in LM relative to EM allows
99 productivity to increase. In the present-day western Arabian Sea, the nutrient enrichment
100 in the surface layer that fuels the seasonal increase in productivity during summer is
101 induced by vertical advection due to upwelling generated by Ekman transport and to wind-
102 driven convective mixing forced by the south-westerly monsoon winds²⁶. The LM
103 simulation depicts atmosphere-ocean interactions in the Arabian Sea that resemble the
104 modern, characterized by strong early summer (JJA) upwelling alongshore Oman and
105 Somalia (locally reaching 1.5 m.day^{-1} upward velocity) and a deepening of the mixed layer
106 (up to 100 m) in the western area (Fig. 2b and 3a and Extended Data Figure 2d). These
107 processes are driven by the south-westerly Somali Jet that generate wind stress off the
108 coasts of Somalia and the Arabian Peninsula (Fig. 3d and 3g). The lack of upwelling in
109 the EM simulation is due to a different Somali Jet strength and pattern with weaker surface
110 and low-level (850 hPa) winds (-6 to -8 m.s^{-1} in the core of the jet compared to LM) broadly
111 shifted equatorward and a more zonal flow between 0°N - 20°N (Fig 3d-e). Consequently,
112 the wind-stress alongshore the Arabian Peninsula decreases (Fig. 3g-i), leading to
113 weaker upwelling (less than 0.6 m.day^{-1} upward velocity) and a flat and shallow (20 m)
114 mixed layer in the whole region (Fig. 3b and Extended Data Figure 2) that prevents
115 nutrient entrainment into surface waters inhibiting productivity. SSTs in the western
116 Arabian Sea are on average 4°C warmer in the EM simulation (Extended Data Figure 2a),
117 because the area receives warm waters from the Neotethyan embayment rather than
118 cooler waters from the equatorial upwelling region advected northwards by the eastern

119 African coastal currents. EM and LM simulations indicate that the onset of seasonally high
120 productivity in the western Arabian Sea as well as SST cooling is caused by a modification
121 of surface ocean circulation, driven by the inception of a Somali Jet similar in structure
122 and intensity to its modern counterpart.

123

124 **Drivers of modern-like SAM winds and rainfall patterns**

125 Further experiments were conducted to assess the mechanisms responsible for the
126 evolution of the Somali Jet and upwelling system during the Miocene. These additional
127 simulations allow us to disentangle the effects of regional paleogeographic evolution and
128 contemporaneous global changes (expanded Antarctic ice-sheet and lower atmospheric
129 CO₂ concentration) on ocean-atmosphere dynamics (Fig. 1, Extended Data Table 1 and
130 Figure 3 and Methods). Major Miocene paleogeographic changes that we considered are
131 the rise of the complex orogen extending from the Anatolian to the Iranian Plateau, the
132 uplift of the East African Highlands and the Himalaya-Tibetan Plateau complex, and the
133 emergence of land in Eastern Arabian Peninsula. Results are synthesized in Figure 4a-
134 b.

135 The most salient outcome of these sensitivity experiments is the key role played by the
136 Middle East physiography in the onset of modern-like atmosphere-ocean dynamics in the
137 Arabian Sea. Removing the Anatolia-Iran orogen (LM-NoAIO vs. LM) significantly
138 reduces wind strength north of 10°N (Extended Data Figure 4c) and weakens upwelling
139 alongshore Oman (Fig. 4a). Additionally submerging the Eastern Arabian Peninsula (LM-

140 NoEAP) in the LM configuration results in the disappearance of the upwelling zone (Fig.
141 3c and 4a). In LM-NoEAP, the Arabian Peninsula immersion suppresses the westward
142 extension of the HTP low pressure anomaly (Extended Data Figure 5). As a result, the
143 model simulates a decrease in south-westerly winds blowing above the western Arabian
144 Sea towards the Indian subcontinent (Fig. 3d,f) and a limited northward expansion of the
145 Somali Jet (Fig 3d,f). Both effects contribute to decreasing the wind stress along the
146 Oman coastlines (Fig. 3g,i), which, in turn, modifies the surface circulation in the western
147 Arabian Sea suppressing upwelling. Meanwhile, expanding the Antarctic ice-sheet (LM-
148 AIS) and reducing atmospheric $p\text{CO}_2$ levels (LM- CO_2) only marginally affect Somali Jet
149 intensity or upwelling (Fig. 4a-b and Extended Data Figure 4a-b), although the imposed
150 $p\text{CO}_2$ decrease results in surface cooling of 1.5°C in the western Arabian Sea (LM- CO_2
151 vs. LM). Cumulated with the $\sim 4^\circ\text{C}$ regional SST cooling induced by the early to late
152 Miocene paleogeographic evolution, our simulations yield a total SST decrease in
153 agreement with proxy reconstructions ($\sim 7^\circ\text{C}$) at nearby drill sites (ODP 722 and 730)¹⁰.
154 Conversely, sensitivity experiments for the early Miocene with modern HTP (EM-HTP),
155 or higher than present-day Himalayan elevation (EM-Him)⁴ show only a marginal impact
156 on atmosphere-ocean interactions in the Arabian Sea (Fig. 4a-b). Sufficiently high East
157 African Highlands help precondition the modern Somali Jet pattern by forcing the cross-
158 equatorial flow to bend (Fig. 4b and Extended Data Figure 4d)¹⁹⁻²⁰. However, changes
159 mainly affect upwelling alongshore Somalia (EM-EAH vs. EM) (Fig 4a) because of
160 restructuring of the jet located in the equatorial area ($0-10^\circ\text{N}$).

161 In addition to its imprint on Arabian Sea surface oceanography, we assess, for each
162 simulation, the intensity of monsoonal circulation through the Webster-Yang Index³⁰,
163 which traces the vigor of atmospheric circulation (Fig. 4c). The intensity of large-scale
164 circulation is preserved in LM simulations regardless of Anatolia-Iran topography or land
165 extent on the Arabian Peninsula. In contrast, all of our EM simulations are characterized
166 by a weaker circulation (Fig. 4c), likely due to the remote control of the large Paratethys
167 Sea on pressure systems^{24,31}.

168 Our results thus corroborate the hypothesis that the western Arabian Sea upwelling
169 system and associated high primary productivity initiated in response to the full
170 development of the "modern-like" Somali jet^{6,11}, which we show was a direct result of the
171 emergence of the Eastern Arabian Peninsula, preconditioned by the East African
172 Highlands and amplified by Anatolia-Iran topography. As the Arabian Peninsula
173 paleogeographic evolution during the Miocene responded both to the long-term Arabia-
174 Eurasia collision³² and to eustatic variations, we hypothesize an indirect contribution of
175 the Antarctica ice-sheet whereby the Mi-3 (13.8 Ma, ~ 50 m) and Mi-4 (12.9 Ma, ~ 20-30
176 m) sea-level retreats³³ increased the emerged surface area of the Eastern Arabian
177 Peninsula and forced abrupt changes¹¹ in regional atmosphere-ocean dynamics.
178 Importantly, our simulations show that the Miocene intensifications of wind strength and
179 upwelling inferred from western Indian Ocean palaeoceanographic records are not
180 directly tied to the evolution of the HTP complex as has often been assumed^{6,7,10,15,16}.

181 Continental rainfall changes are temporally decoupled from Arabian Sea upwelling
182 changes. Although the full emergence of the Arabian Peninsula in the LM is instrumental

183 in driving the onset of Oman upwelling, it does not markedly alter either of the precipitation
184 metrics (Fig 4a,d-e and Extended Data Figure 6 and 7) because moisture transport by
185 the Somali Jet to peninsular India only slightly increases (Extended Data Figure 8f). In
186 contrast, the Anatolia-Iran uplift enhances both upwelling and summer precipitation on
187 Himalayan topography and South Asian lowlands (Fig 4d-e and Extended Data Figure 6
188 and 7). Anatolia-Iran Plateau uplift deepens the low-pressure area over the Arabian
189 Peninsula, enhancing low-level wind intensity and moisture transport from the Arabian
190 Sea toward the South-Asian landmass (Extended Data Figure 8e). EM, increasing
191 topography in the HTP region alone (EM-HTP and EM-Him) increases summer
192 precipitation and seasonality over the region's highlands (Fig. 4d and Extended Data
193 Figure 6-7), while the Somali Jet and upwelling intensity are only marginally affected (Fig
194 4a-b). These results corroborate the hypothesis that the increases seen in sedimentary
195 and weathering fluxes between 25 and 15 Ma could be partly associated with monsoon
196 intensification related to coeval uplift of the Himalayan ranges⁵, although weathering rates
197 would also be sensitive to change in exhumation rates recorded at that time.

198

199 **Reconciling divergent South Asian Monsoon records**

200 The history of the SAM remains hotly debated as new paleoceanographic and continental
201 records emerge and propose contradictory timing for the onset of modern-like Asian
202 monsoons^{3,6,11,31,34-37}. Our new set of climate simulations shows that these apparent
203 contradictions can be reconciled. From our results we infer that seasonally intense
204 precipitation in South Asia existed in the early Miocene and probably earlier, as suggested

205 by continental records and paleoclimate models^{3,31,37}, but that a modern-like structure and
206 vigor of the SAM atmospheric circulation developed in the late middle Miocene as a result
207 of regional changes in Middle-East and East-Africa, in agreement with multi-proxy
208 palaeoceanographic records^{6,7,10,11}. This polygenetic history of the SAM is coherent with
209 the proposed Neogene transition from an 'ITCZ-dominated' to a 'land-sea breeze-
210 dominated' monsoon^{4,37} based on botanical evidence. Here, we show that this transition
211 towards the modern monsoon system is independent of Asian orography but is forced by
212 geographical evolution at the western boundary of the Indian Ocean.

213 This emerging view does not preclude an important role for the Himalaya-Tibet orogeny
214 in strengthening rainfall amount and in the establishment of a longer rainy season (similar
215 to modern) in the early Miocene or before^{5,13,38}. However, our results do indicate that HTP
216 tectonic activity cannot be held to account for either the palaeoceanographic changes
217 observed in the Arabian Sea and equatorial Indian Ocean, or for the establishment of
218 modern-like large-scale atmospheric circulation with a strong Somali Jet. Our results
219 further emphasize the important role of the Iran-Zagros orogen on monsoonal rainfall in
220 agreement with ref.17. Hence, while much of the effort to understand the evolution of
221 Cenozoic Asian monsoons has focused on the Himalaya-Tibet region, we underscore that
222 constraining the exact timing of East African and Middle Eastern physiographic
223 changes is crucial to grasp the full complexity of Asian monsoon evolution. Determining
224 the timing of the initiation of a "true" modern SAM is a semantic issue, the answer to which
225 depends on which metric primarily defines the modern SAM: the atmospheric circulation
226 and regional atmosphere-ocean dynamics, the rainfall intensity, or the seasonal

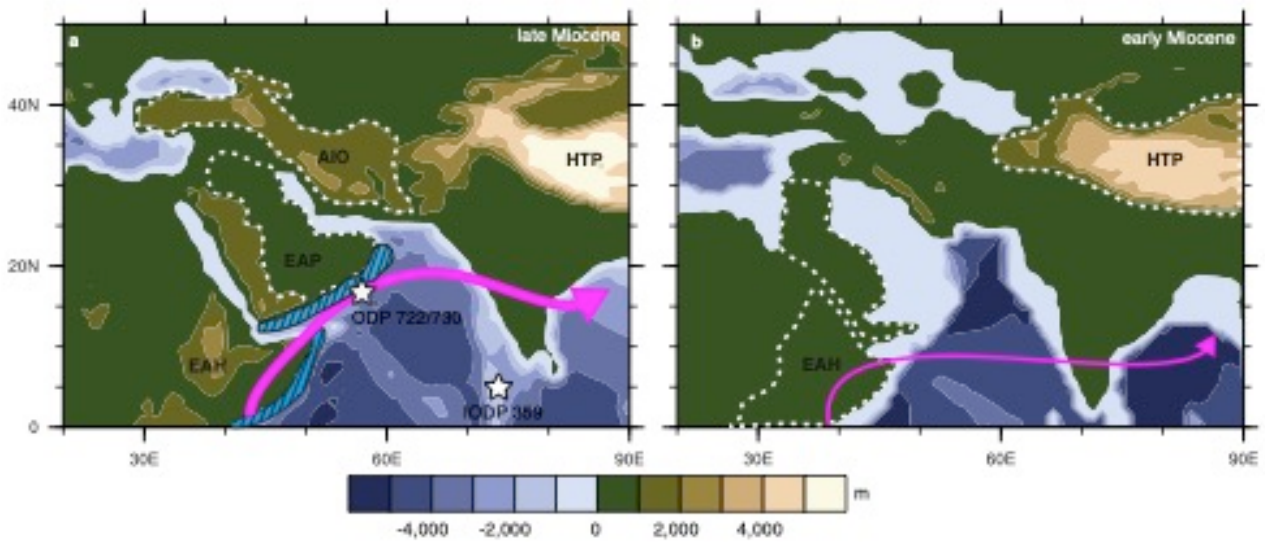
227 distribution of rainfall. We have shown here that, on geological timescales, these three
228 parameters are not controlled by the same factors and likely had independent histories.

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236 to deep-time configurations.

237 **Author contributions:** A-CS and YD designed the study and ran the simulations. FF
238 provides updated paleogeographies and expertise on paleogeography evolution. JBL and
239 ML developed and ran the tests for the run-off adapted version of PISCES and help with
240 the setup of PISCES simulations. CTB helped compile and synthesize paleoceanographic
241 records. A-CS, YD, CTB, AL and JBL wrote the manuscript. GDN, FF and DT provided
242 substantial comments and revisions.

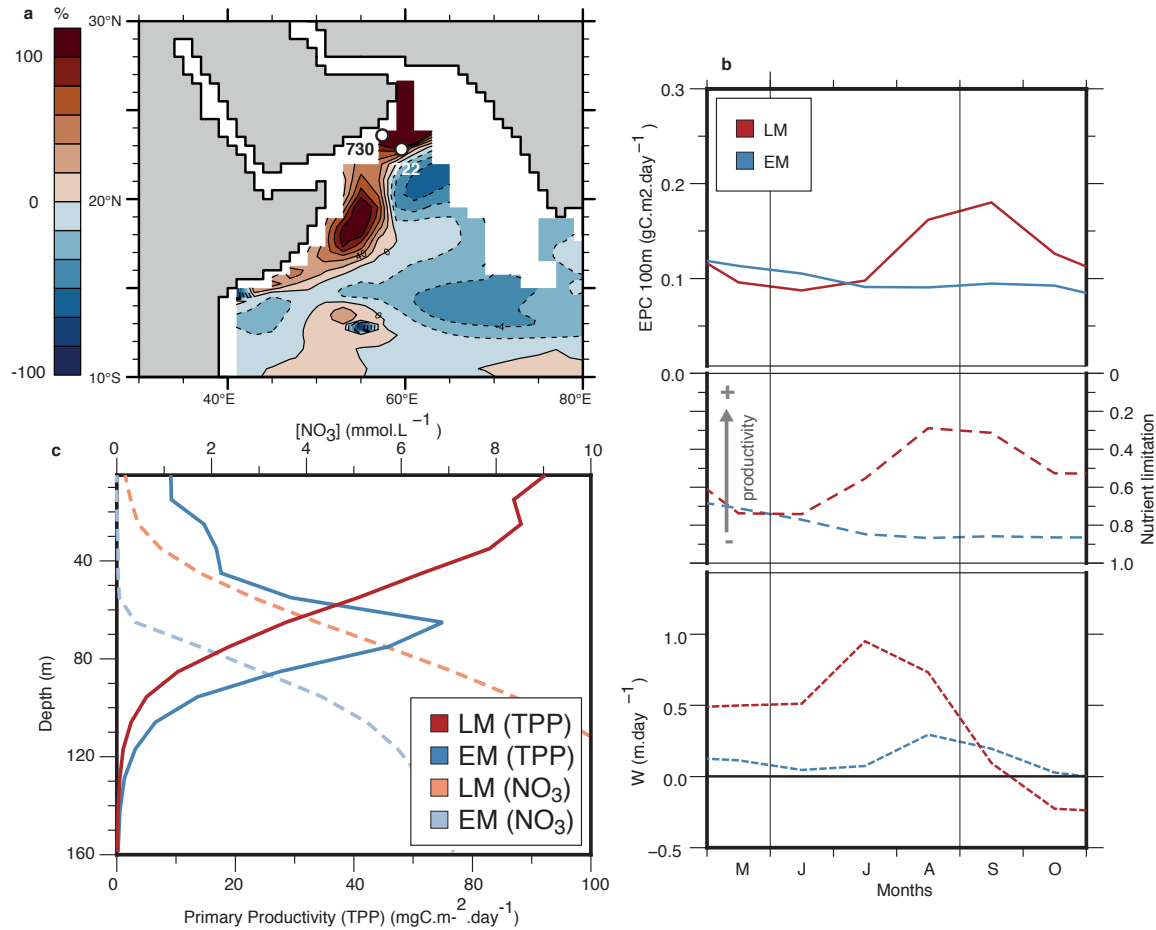
243 **Competing interests:** Authors declare no competing interests.

244 **Figure captions:**



245 **Fig. 1. Western Indian Ocean paleogeographic reconstructions.** (a) late Miocene
 246 (~10 Ma) (this study, updated from ref. 27, see Methods) ; (b) early Miocene (~20 Ma)²⁸.
 247 EAH - East African Highlands; AIO – Anatolian-Iran Orogen; HTP - Himalaya-Tibetan
 248 Plateau; EAP - East Arabian Peninsula. Present-day geography and ocean-atmosphere
 249 dynamics resemble the late Miocene in this region. Dashed contours show areas where
 250 modifications are applied in sensitivity experiments (see Methods and Extended Data
 251 Table 1 and Figure 3 for detailed descriptions). Simulated Somali Jet patterns and
 252 strengths are represented by the magenta arrows. Blue stretched areas show upwelling
 253 location from simulations. Stars show the location of drilling sites or expeditions cited in
 254 the text. IODP - International Ocean Discovery Program; ODP - Ocean Drilling Program.

255



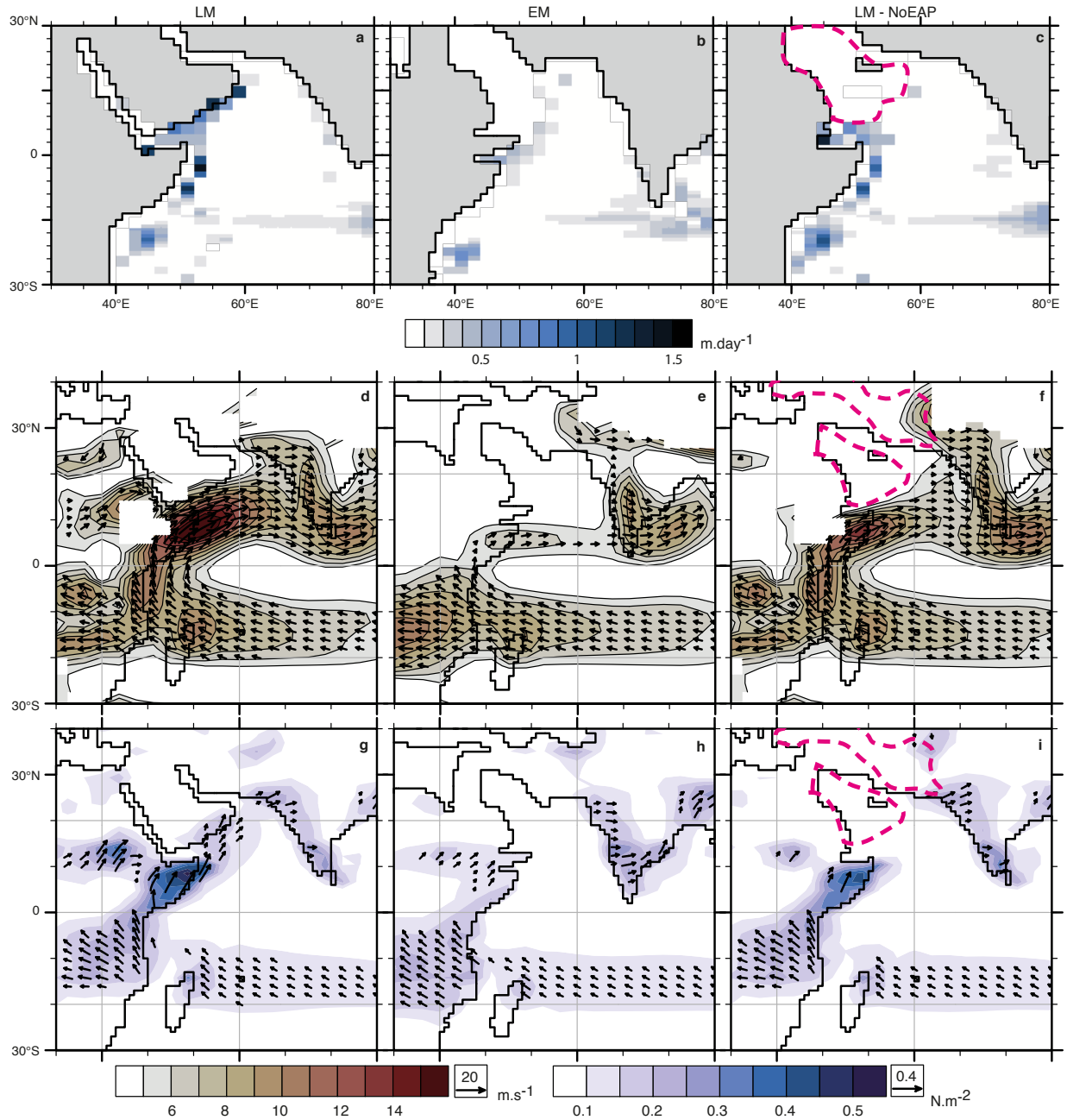
256 **Fig. 2. Change in Arabian Sea productivity between the early and late Miocene.** (a)
 257 Change in Export Production at 100m (EPC 100m) between early Miocene (EM) and late
 258 Miocene (LM) simulations during late summer (August-October). ODP Sites 722 and 730
 259 are shown on the map; (b) Seasonal variation of EPC 100m along the Oman Margin
 260 (gC.m².day⁻¹) (averaged along coastal grid points north of 15°N), nutrient limitation
 261 (averaged over 0-60 m, note inverted axis), and mean vertical velocity (W) at 80 m (m.day⁻¹).
 262 A limitation term of zero indicates no nutrient limitation. In panels a and b, Export
 263 Production at 100 m (EPC 100m), a commonly used measure of the carbon export to the
 264 deep ocean that is ultimately recorded in sediments, is shown instead of Primary
 265 Productivity but the two exhibit similar temporal behavior; (c) July-October cumulated

266

Total Primary Productivity (TPP, $\text{mgC}\cdot\text{m}^{-2}\cdot\text{day}^{-1}$, solid lines) and averaged Nitrate

267

concentration (NO_3 , $\text{mmol}\cdot\text{L}^{-1}$, dashed lines) with depth for EM and LM.



268

269

Fig. 3. Changes in ocean-atmosphere dynamics in response to Miocene

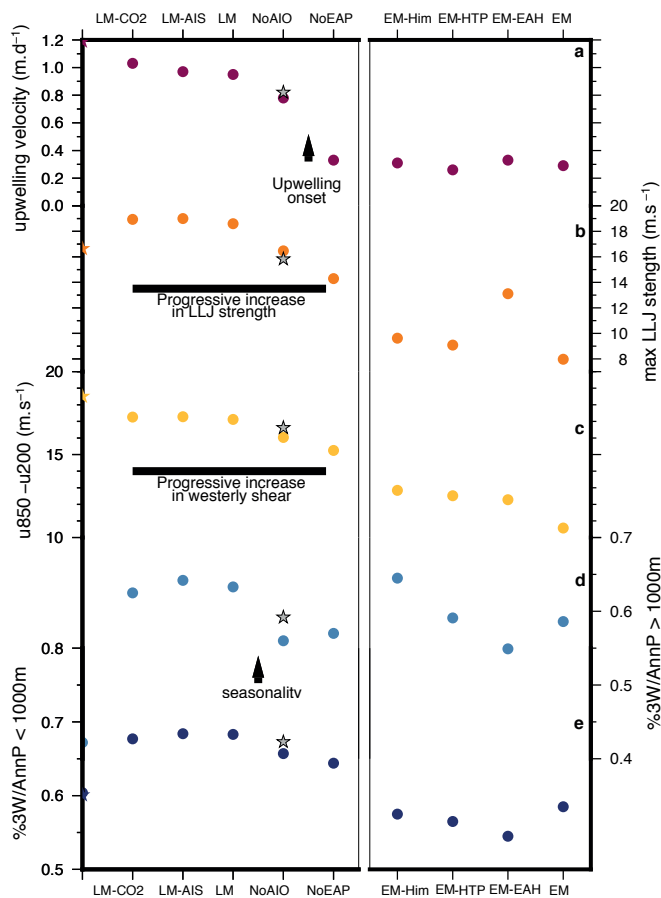
270

paleogeographic evolution. Top: Upwelling velocity (vertical velocity at 80 m depth,

271

averaged over JJA); Middle: Low level winds (850 hPa) during boreal summer (JJA);

272 Bottom: Wind stress during boreal summer (JJA). (a), (d) and (g) - late Miocene (LM) and
 273 (b), (e) and (h) – early Miocene (EM) simulations; (c), (f) and (i) LM-NoEAP. LM-NoEAP
 274 is a simulation without the Eastern Arabian Peninsula (EAP), designed to show the
 275 influence of Arabian Peninsula immersion on the Somali Jet structure in a LM
 276 configuration. Dashed pink contours on panels (c), (f) and (i) indicate the location of
 277 geographic modifications compared to LM simulation (panels (a), (d) and (g)).



278
 279 **Fig. 4. South Asian Monsoon circulation and rainfall in sensitivity experiments.** (a)
 280 Maximum summer (JJA) vertical velocity at 80 m averaged over coastal grid points north
 281 of 15°N ; (b) maximum intensity of the Somali Jet over the Arabian Sea (averaged over
 282 JJA, [30-60°E, 0-20°N]); (c) Webster-Yang Index (Meridional wind stress shear

283 (u850hPa-u200hPa) averaged over [40-110°E, 0-20°N] and June-September)³⁰ ; (d) 3
284 consecutive wettest months relative to annual mean precipitation [65-95°E, 0-35°N, see
285 Extended Data Figure 6] for high-elevation areas (over 1,000 m - light blue) and e) for
286 Indian foreland (below 1,000 m - dark blue). A 1,000 m threshold was chosen to
287 distinguish low-land areas of the Indian subcontinent from high topography of the
288 Himalaya-Tibetan Plateau (HTP) at the model grid resolution. Seasonality changes are
289 mostly driven by changes in summer rainfall amount (Extended Data Figure 6 and 7).
290 Colored stars indicate values for a pre-industrial simulation²⁵. Gray star indicates LM-
291 NoEAHR simulation which account for uncertainties in late Miocene East African Highland
292 (EAH) elevation (see “Methods”). Simulation characteristics can be found in Extended
293 Data Table 1 and Extended Data Figure 1 and 3.

294

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390
391

392 **Methods**

393

394 **Model General Description**

395 The IPSL-CM5A2²⁵ Earth System Model (ESM) is an updated version of the IPSL-CM5A-
396 LR model ESM³⁹. This version benefits from new numerical developments enhancing
397 computational performance, which makes it suitable for deep-time paleoclimate
398 simulations. The model is composed of the atmosphere model LMDz5A⁴⁰, the land
399 surface and vegetation model ORCHIDEE⁴¹ and the ocean model NEMO_v3.6⁴² that
400 includes an ocean dynamics model (OPA), a thermodynamic-dynamic sea-ice model
401 (LIM2)⁴³ and a biogeochemistry model (PISCES-v2)²⁶. The ocean model grid has a
402 nominal horizontal resolution of 2° by 2° refined up to 0.5° in the equatorial region and 31
403 vertical levels, whose thickness varies from 10 m near the surface to 500 m at the bottom.
404 The atmosphere grid has a nominal horizontal resolution of 3.75° in longitude by 1.9° in
405 latitude with 39 irregularly distributed vertical levels. Model components are fully
406 described in refs. 25,39. IPSL-CM5A2 has been used for several paleoclimatic
407 studies^{44,45}, including work focused on the Asian Monsoons³¹, and is part of the deep-
408 time Model Intercomparison Projects focused on the Pliocene (PlioMIP2)²⁷, the mid-
409 Miocene (MioMIP)⁴⁶ and the early Eocene (DeepMIP)⁴⁷. Detailed evaluation of the model
410 performance at simulating modern Asian monsoons can be found in ref. 31. General
411 atmospheric circulation over the Asian region and specific characteristics of the South
412 Asian monsoon are correctly simulated by IPSL-CM5A2. Room for improvement exists in
413 the temporality and intensity of the South Asian monsoon as the monsoon onset lags by

414 one month compare to observation and the rainfall intensity during summer over low
415 elevation area is slightly underestimated (Extended Data Figure 7a).

416 **Adaptation of the PISCES model for deep-time studies**

417 The PISCES model simulates the biogeochemical cycles of carbon and main nutrients
418 and lower trophic levels of the marine ecosystem²⁶. It includes the representation of 2
419 types of phytoplankton (nanophytoplankton and diatoms) and 2 types of zooplankton
420 functional types and five limiting nutrients (phosphate, nitrate, iron, ammonium and
421 silicate). Phytoplankton growth is limited by the availability of nutrients and light, and water
422 temperature. PISCES also simulates two size classes of particulate organic carbon (small
423 and large) that differ by their sinking rates, as well as the semi-labile dissolved organic
424 matter, dissolved inorganic carbon, alkalinity and dissolved oxygen. A complete
425 description of model parameterizations and evaluation is found in ref. 26. PISCES has
426 been widely used for studying the relationship between marine productivity and global
427 climate^{48,49}, including paleoclimates⁵⁰⁻⁵³. In particular, PISCES has also been employed
428 to investigate the relationship between the Indian Monsoon and productivity in the
429 Western Arabian Sea during the Quaternary⁵³ as well as the impact of ocean physics and
430 dynamics on productivity in the tropical Indian ocean²⁹. The overall productivity in the
431 Arabian Sea is correctly simulated⁵⁴ though with weaker values compared to observations
432 owing to the weaker Somali jet, which in turn, impacts the coastal upwelling extent and
433 intensity. The absence of meso-scale dynamics due to the moderate ocean resolution
434 also explains some differences between modeled productivity patterns and
435 observations⁵⁴. Despite those well-known biases, the model depicts a realistic

436 representation of the physical mechanisms driving productivity, and the seasonality of
437 productivity blooms is well simulated⁵³.

438 For this study, we have updated the scheme employed to compute the river supply of
439 nutrients and other elements because the paleogeography is significantly different from
440 present-day⁵⁵. In the original version, elements' delivery is fixed and uses results from the
441 Global Erosion Model⁵⁶ for DIC and alkalinity or is taken from the GLOBAL-NEWS2 data
442 sets⁵⁷ for other elements. Here, the riverine nutrient input to the ocean is calculated as
443 the simulated model runoff times the riverine concentration for each element. As such,
444 the element supply to the ocean is consistent with the Miocene paleogeography and
445 simulated continental runoff. We keep the riverine concentration in elements across the
446 globe constant because there is no simple way of determining how element
447 concentrations might have varied according to local soil composition, vegetation and
448 climate. More importantly, this concentration is adjusted so that the total annual global
449 amount in each supplied element is conserved between the fixed modern supply and the
450 runoff-dependent supply. By doing that, our simulations are designed to isolate and to
451 quantify the effect of ocean dynamics on primary productivity.

452 **Baseline boundary conditions**

453 Late and early Miocene paleogeographies are shown in Extended Data Figure 1. The late
454 Miocene paleogeography is based on the PRISM4²⁷ reconstruction used in PlioMIP2 ⁵⁸
455 (available on the following link : https://geology.er.ugs.gov/egpsc/prism/4_data.html) with
456 additional manual corrections.

457 The position of the continents is therefore close to present-day. The elevation of
458 mountains belts (including HTP region) is similar to present-day. Major differences include
459 the removal of Hudson Bay and the closure of Bering Strait²⁷. The Australian continent is
460 shifted southward to account for its northward migration throughout the Cenozoic⁵⁹ and
461 the Sunda shelf has been partly emerged⁶⁰. The early Miocene paleogeography is taken
462 from the recent study of Poblete et al. (2020)²⁸. Most of the mountain belts have lower
463 elevation compared to the late Miocene paleogeography to account for major phases of
464 uplift recorded during the late Neogene. In addition, the tip of India is located closer to the
465 equator. The Neo-Tethys seaway is kept closed by a land-bridge (Gomphotherium
466 landbridge⁶¹) so there is no water mass exchange between Indian and Atlantic Oceans
467 at low latitudes in the early Miocene paleogeography. Another major feature is the
468 existence of a large Peri-Tethys Sea, which covers a substantial part of the European
469 and Asian continents⁶¹. In the absence of global vegetation reconstructions for both the
470 early and the late Miocene, we use a conservative approach by imposing idealized
471 vegetation with a latitudinal distribution in the two configurations as already done in refs.
472 31,44. Though vegetation may potentially alter local atmospheric dynamics, the latest
473 research suggests that the large-scale pattern of atmosphere-ocean circulation in the
474 tropical Indian Ocean is mostly driven by SSTs gradients¹⁷. In order to tease apart the
475 effect of paleogeography alone, the simulations are run with a prescribed atmospheric
476 $p\text{CO}_2$ concentration of 560 ppm whereas other greenhouse gases are set at their pre-
477 industrial values. The solar constant is set at $1364.3 \text{ W}\cdot\text{m}^{-2}$ for LM and $1362.92 \text{ W}\cdot\text{m}^{-2}$
478 for EM and orbital parameters are kept at modern values.

479

480 **Sensitivity experiments**

481 In addition to simulations performed with baseline configurations (i.e. EM and LM), we
482 investigate the sensitivity of the Arabian Sea atmosphere-ocean dynamics and of the
483 South Asian monsoon patterns to regional changes in topography. These sensitivity
484 experiments consist of altering the topography of either the Anatolian-Iranian Orogen
485 (AIO), the Eastern Arabian Peninsula (EAP), the East African Highlands (EAH) or the
486 Himalayan Tibetan Plateau (HTP), and are either performed with the late or early Miocene
487 baseline paleogeographies (Extended Data Table 1 and Extended Data Figure 3). We
488 focused on the major regional changes in topography and land-sea distribution around
489 the Indian Ocean basin that have been suggested in the literature to drive first-order
490 changes in large-scale monsoon dynamics. These sensitivity experiments integrate
491 existing uncertainties in paleogeographic reconstructions because the intricate uplift
492 history of Asian, East African and Middle-East orogens is still a subject of active research.

493 Thermochronology and paleoaltimetry studies indicate that the Central Tibetan
494 Plateau had reached high elevations (> 4,000 m) by the early Miocene^{12,62-66}. The
495 Himalayan mountain range attained high elevations similar to present day in the late early
496 Miocene (~ 15 Ma)^{12,13,66}, a configuration we test in the EM-HTP configuration (Extended
497 Data Figure 3g). Some studies also suggest a configuration in which the Himalayan
498 topography is higher than today, possibly sustained by slab break off, to explain rainfall
499 intensification during the early to middle Miocene⁶⁶. We test this scenario in the EM-Him
500 simulation (Extended Data Figure 3h).

501 Uplift in East Africa initiated in the Eocene/Oligocene with the doming in Ethiopia
502 and Kenya⁶⁷. This episode corresponds to the establishment of long-wavelength
503 topography associated with mantle dynamics (see ref⁶⁸ for a review). In Ethiopia, traps
504 volcanism occurred during the Early Oligocene⁶⁸ leading to the formation of ~1km thick
505 continental flood basalts covering pre-existing topography. The Ethiopian plateau can be
506 possibly as high as 2500-3000 m in place before the middle Miocene⁶⁹⁻⁷². East African
507 Highlands (East African Rift and Afar-Yemen-Arabia Plateau) then underwent continuous
508 uplift during the Neogene^{69,72} owing to rifting processes : beginning in Ethiopia during the
509 middle Miocene and then propagating southwards⁶⁸. Topography in the central Kenya
510 region was probably high in the middle Miocene (above 1,400m elevation before 13.5
511 Ma)⁷². We test the effect of uplifting topography in East Africa in the EM-EAH
512 configuration, in which we prescribed the LM EAH topography onto the EM baseline
513 geography (Extended Data Table 1 and Extended Data Figure 3f). The impact of late
514 Miocene uplift of rift-associated topography has also been tested on the LM-NoAIO
515 geography by capping the elevation at 1,500 meters (an altitude likely reached before the
516 middle Miocene) but maintaining the morphology of the large-scale topography (Extended
517 Data Figure 3d ; LM-EAHR). This latter sensitivity experiment have results similar to LM-
518 NoAIO, which suggest that late Miocene uplift related the EAH rift activity have had little
519 impact on the atmospheric circulation and upwelling. Results on Figure 4d-e however
520 shows that rainfall in the SAM region are sensitive to African topography²⁰⁻²¹ a question
521 that should be further explored.

522 In the Middle-East, the Iranian Plateau and Anatolian topography rise at some
523 point after 17 Ma due to the long-term collision of the Arabian and Eurasian Plates³², most
524 likely during the late Miocene^{73,74}, which is later than the settlement of high topography in
525 East-Africa⁶⁷⁻⁷² and HTP regions^{12,62-66}. We therefore run an experiment in which we
526 decrease the height of regional topography by half on the LM configuration (LM-NoAIO,
527 Extended Figure 3b).

528 The Tethyan Seaway also closed permanently around 14 Ma^{61,75} and continental
529 area replaces marine environments in the Eastern Arabian Peninsula⁶¹ due both to
530 tectonics evolution of the Middle East³² and to the sea level drop following Antarctic ice-
531 sheet expansion during the Middle Miocene Climate Transition (up to 80 m³³). Existing
532 paleogeographic reconstructions shows the Eastern Arabian Peninsula as partly
533 submerged during the Burdigalian (early Miocene) and emerged in the late Miocene⁶¹.
534 Because of its flatness, the north-eastern part of Arabian Peninsula is highly sensitive to
535 sea level fall and we hypothesize that the sea level drop during the MMCT could have led
536 to emergence of previously submerged land surface. Based on this hypothesis, the effect
537 of changes in land-sea extension over the Eastern Arabian Peninsula is tested in the LM-
538 NoEAP configuration (Extended Data Figure 3c). Other details of regional
539 paleogeography, such as narrow mountain belts (e.g. the Western Ghats in India)) have
540 not been considered in the present study. Although these small-scale features influence
541 rainfall and wind patterns at the local scale^{76,77} they do not represent first order controls
542 on the large-scale, trans-oceanic, monsoon dynamics¹⁷.

543 In addition to sensitivity experiments on topography and land-sea distribution, we
544 also assess the impact of coeval global climatic changes occurring during the late middle
545 Miocene using sensitivity simulations with an expanded Antarctic ice-sheet^{78,79} or a lower
546 CO₂ concentration⁸⁰ (Extended Data Table 1).

547

548 **Experimental design**

549 IPSLCM5A2 simulations are initialized with idealized ocean conditions (except EM-Him
550 and EM-HTP that restart from EM, Extended Data Table 1) consisting of a latitudinally-
551 varying, zonally symmetric, temperature distribution and a constant salinity distribution⁸¹

$$552 \quad T(^{\circ}\text{C}) = (1000 - z)/1000 * 20 \cos(\text{lat}) + 10 \text{ for } z \leq 1000 \text{ m}$$

$$553 \quad T(^{\circ}\text{C}) = 10 \text{ for } z > 1000 \text{ m}$$

$$554 \quad S \text{ (PSU)} = 34.7$$

555 Each simulation is run for more than 2,500 years, until the deep ocean reaches a quasi-
556 equilibrium with only small residual trends of less than 0.1°C per century (Extended Data
557 Figure 9). Our simulations are analyzed and discussed using climatological averages
558 calculated over the last 100 model years. In a second step, we use the climatology
559 corresponding to each IPSLCM5A2 simulation to force the offline version of PISCES in
560 order to increase the spin-up time of marine biogeochemistry⁸². In each offline PISCES
561 simulation, the prescribed nutrient concentration in rivers is adjusted to the total runoff
562 flux to conserve the total modern global amount of nutrients delivered to the ocean. The

563 global amount of nutrients is thus fixed to its modern value, which enable us to attribute
564 modifications in primary productivity to changes in ocean dynamics.

565 **Model evaluation**

566 We provide here a basic assessment of the model performance compared to available
567 Miocene SST estimates (Extended Data Figure 9b-c, compilation from ref. 45 and
568 reference therein) and note that a more detailed investigation can be found in ref. 45 that
569 is part of the MioMIP project. The model-data comparison shows an overall good fit for
570 both EM and LM simulations (Extended Data Figure 9b-c), in particular in the tropical to
571 mid-latitude domains, where our study is focused (northern Indian Ocean and South
572 Asia). There are some discrepancies at higher latitudes in the Atlantic Ocean that are
573 systematic characteristics of multi-model Miocene simulations because models fail to
574 reproduce the polar amplification and reduced equator-pole SST gradient inferred from
575 the data record⁴⁶.

576 **Data Availability**

577 All model outputs used in this study are available as NetCDF file on the following Zenodo
578 repository⁸⁵ : <https://doi.org/10.5281/zenodo.5727042>. The early Miocene
579 paleogeographic reconstructions²⁸ is also available on the Paleoenvironment map
580 website (<https://map.paleoenvironment.eu>). The repository also contains paleogeography
581 grids used for the simulations. Colored figures in this paper were made with perceptually
582 uniform, color-vision-deficiency-friendly scientific color maps, developed and distributed
583 by Fabio Crameri⁸⁶

584

585 **Code availability**

586 LMDZ, XIOS, NEMO and ORCHIDEE are released under the terms of the CeCILL
587 license. OASIS-MCT is released under the terms of the Lesser GNU General Public
588 License (LGPL). IPSL-CM5A2 source code is publicly available through svn, with the
589 following commands line :svn co
590 http://forge.ipsl.jussieu.fr/igcmg/svn/modipsl/branches/publications/IPSLCM5A2.1_1119
591 2019 modipsl ; cd modipsl/util ; ./model IPSLCM5A2.1

592 The mod.def file provides information regarding the different revisions used, namely : -
593 NEMOGCM branch nemo_v3_6_STABLE revision 6665
594 - XIOS2 branches/xios-2.5 revision 1763
595 - IOIPSL/src svn tags/v2_2_2
596 - LMDZ5 branches/IPSLCM5A2.1 rev 3591
597 - branches/publications/ORCHIDEE_IPSLCM5A2.1.r5307 rev 6336 - OASIS3-MCT
598 2.0_branch (rev 4775 IPSL server)

599 The login/password combination requested at first use to download the ORCHIDEE
600 component is anonymous/anonymous. We recommend that you refer to the project
601 website:

602 http://forge.ipsl.jussieu.fr/igcmg_doc/wiki/Doc/Config/IPSLCM5A2 (last access: 7
603 February, 2022) for a proper installation and compilation of the environment.

604 Adaptation of PISCES model used in the study have been archived on the following
605 Zenodo repository : 10.5281/zenodo.5727042) in addition with information on how to
606 include the updates in the reference code of PISCES.

607 Analysis and graphics from this paper have been done using open source tools. PyFerret
608 is a product of NOAA's Pacific Marine Environmental Laboratory (information is available
609 at <http://ferret.pmel.noaa.gov/Ferret/>). Information on NCL⁸⁷ is available at
610 <https://www.ncl.ucar.edu>. Information on Generic Mapping Tool⁸⁸ is available at
611 <https://www.generic-mapping-tools.org>.

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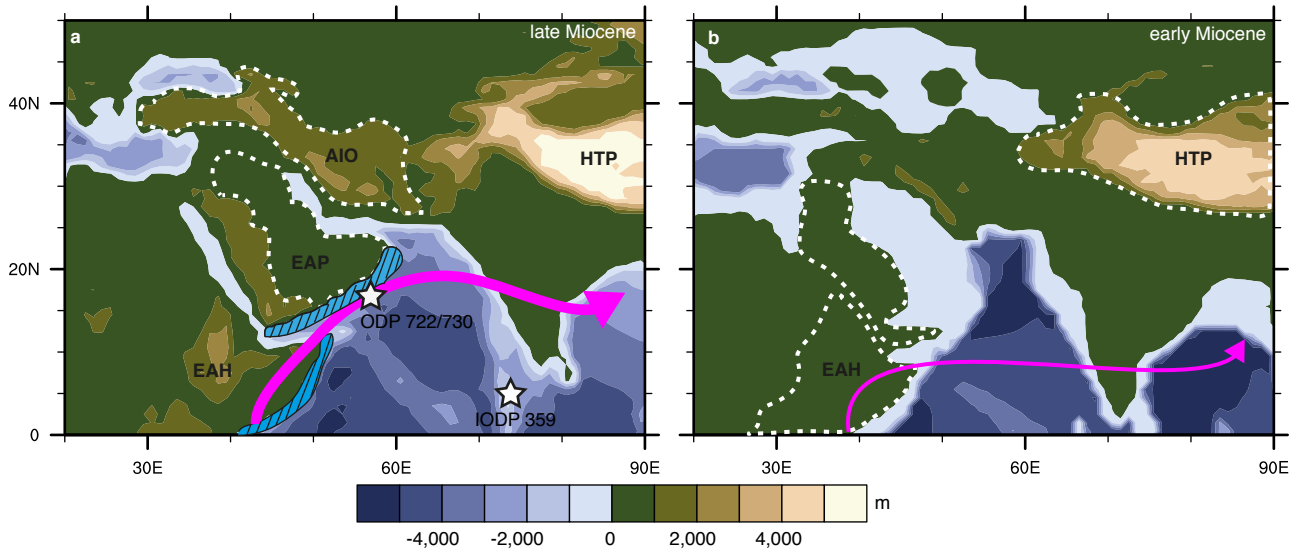
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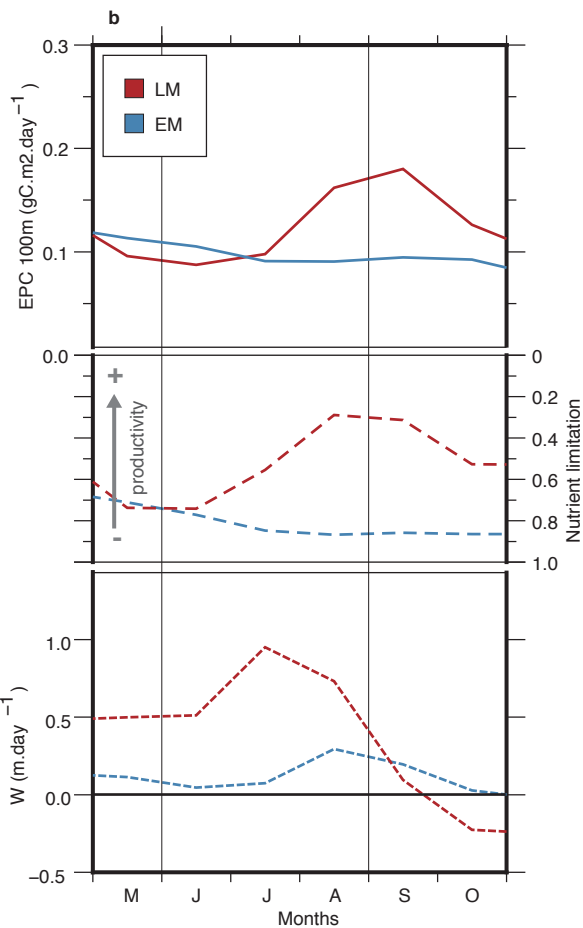
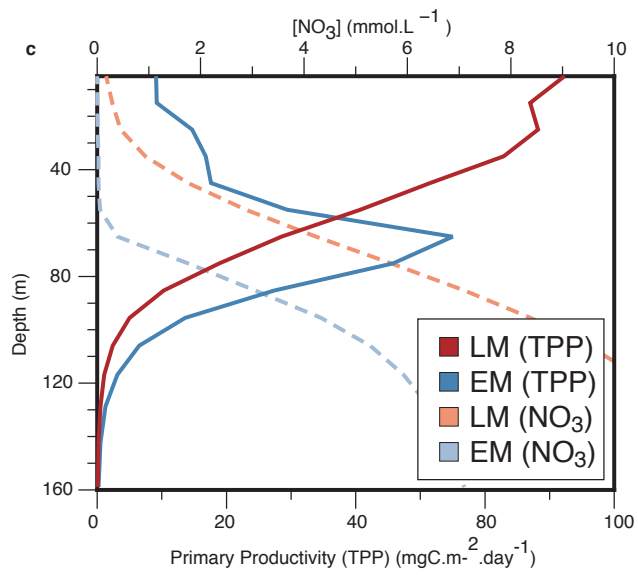
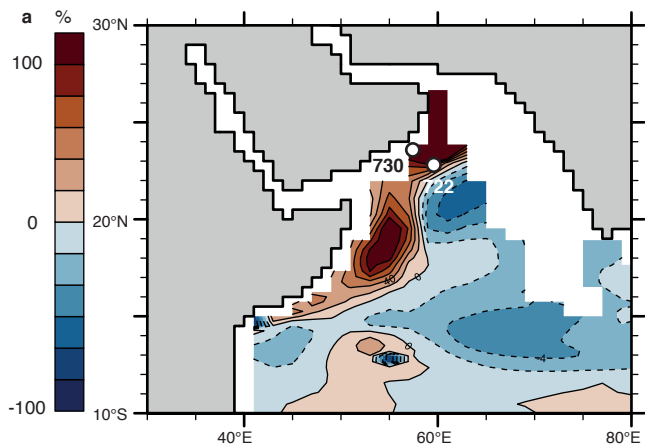
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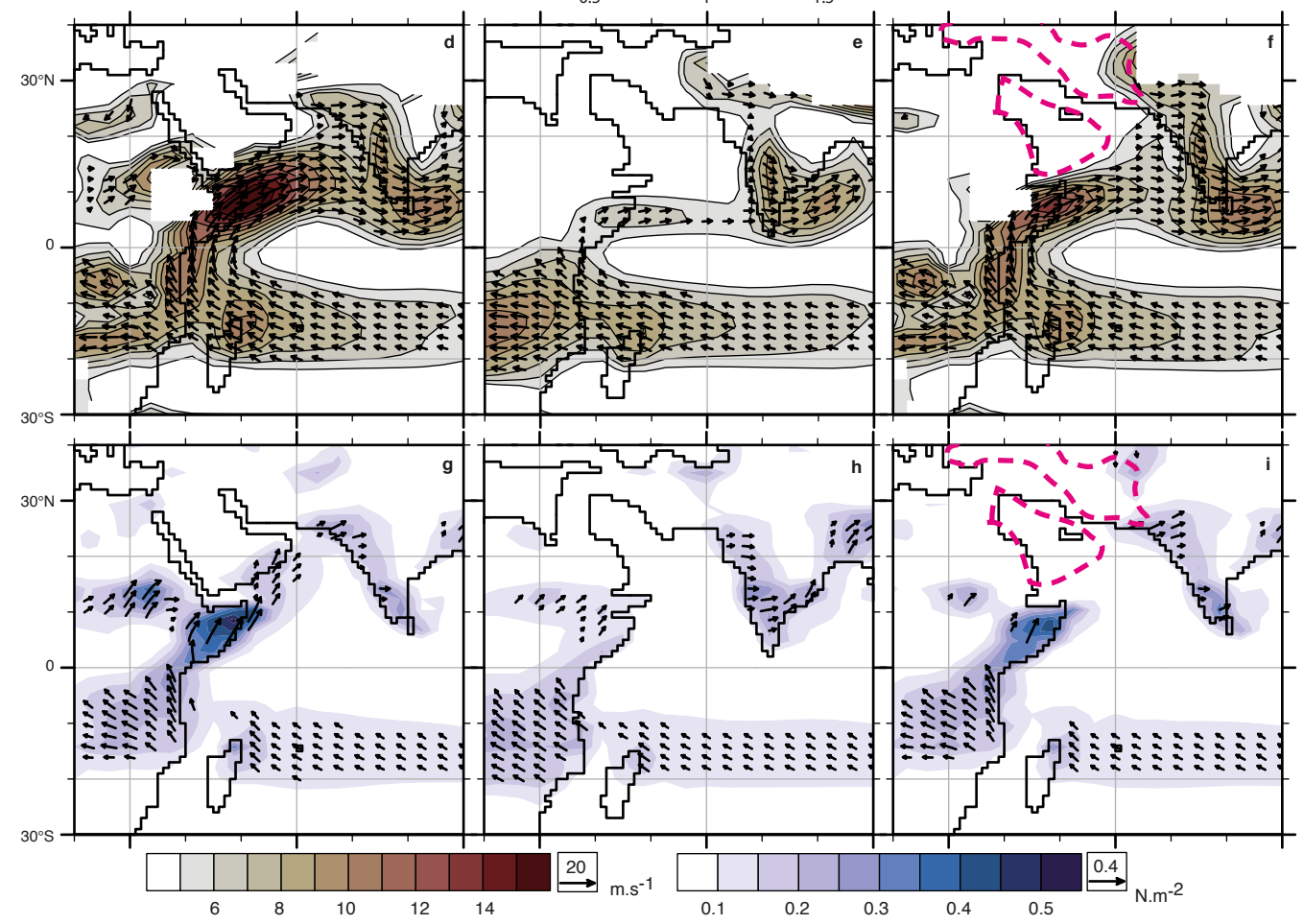
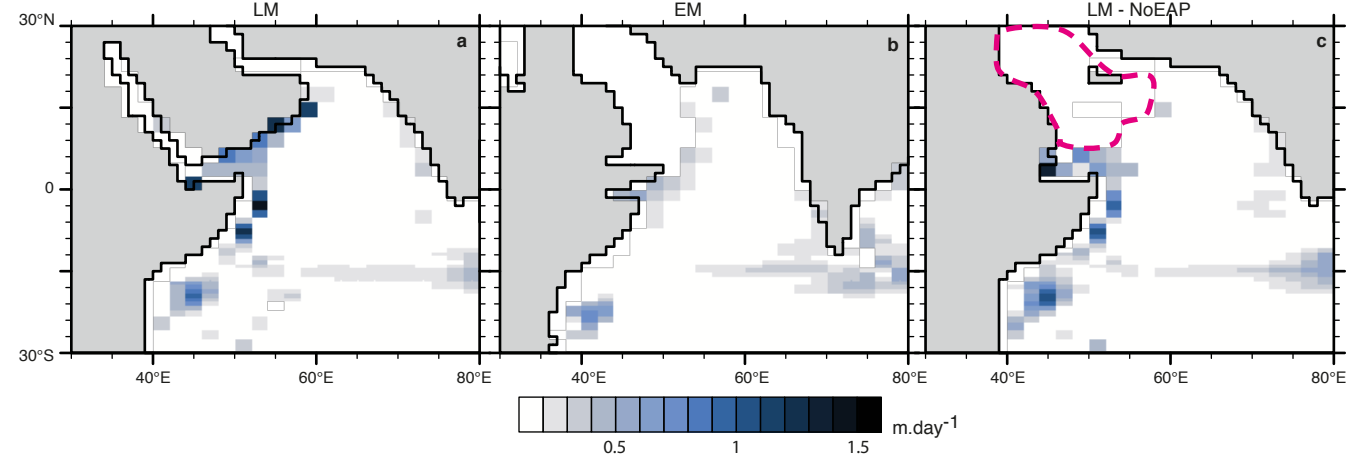
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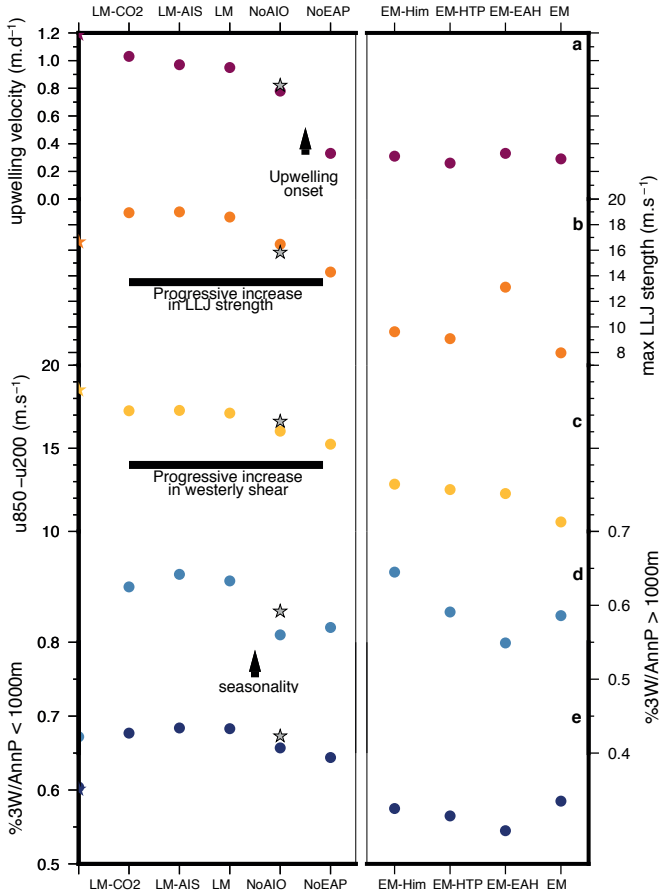
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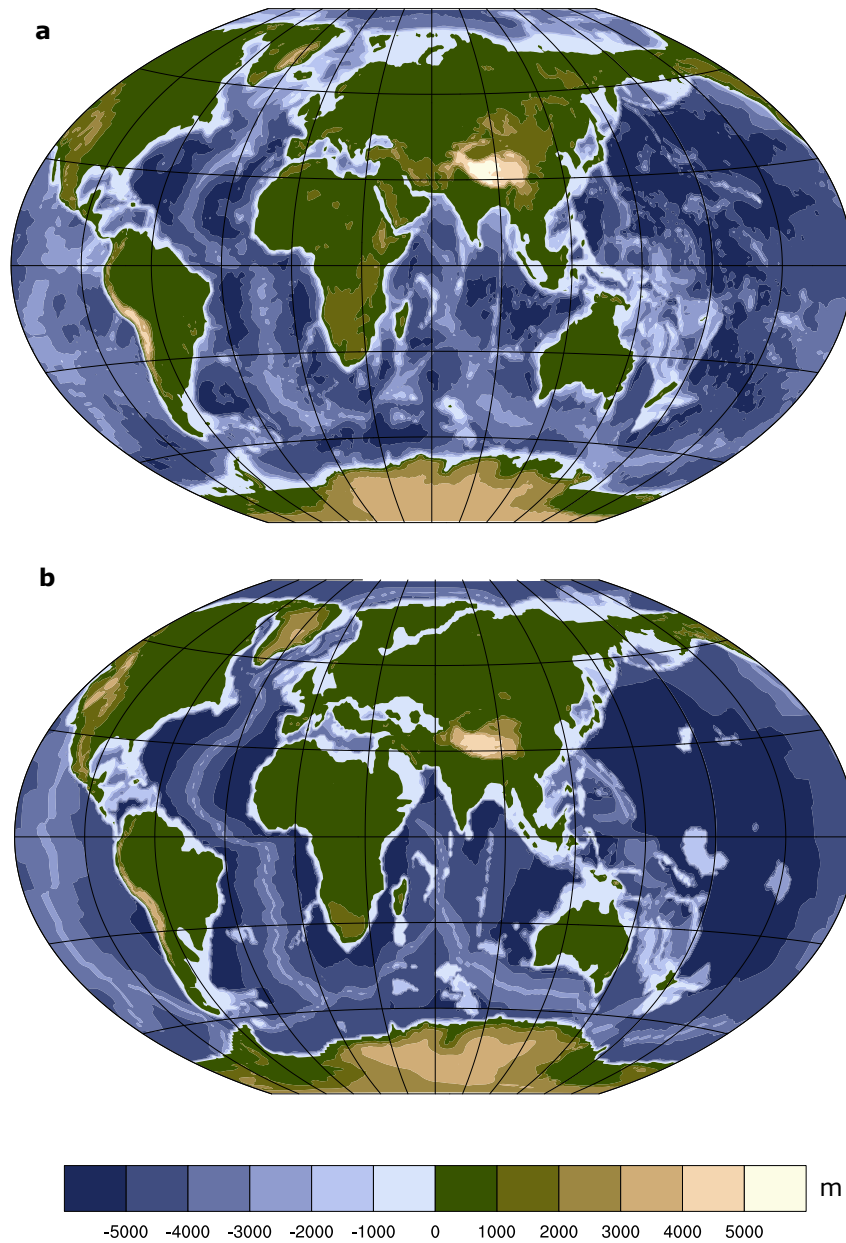




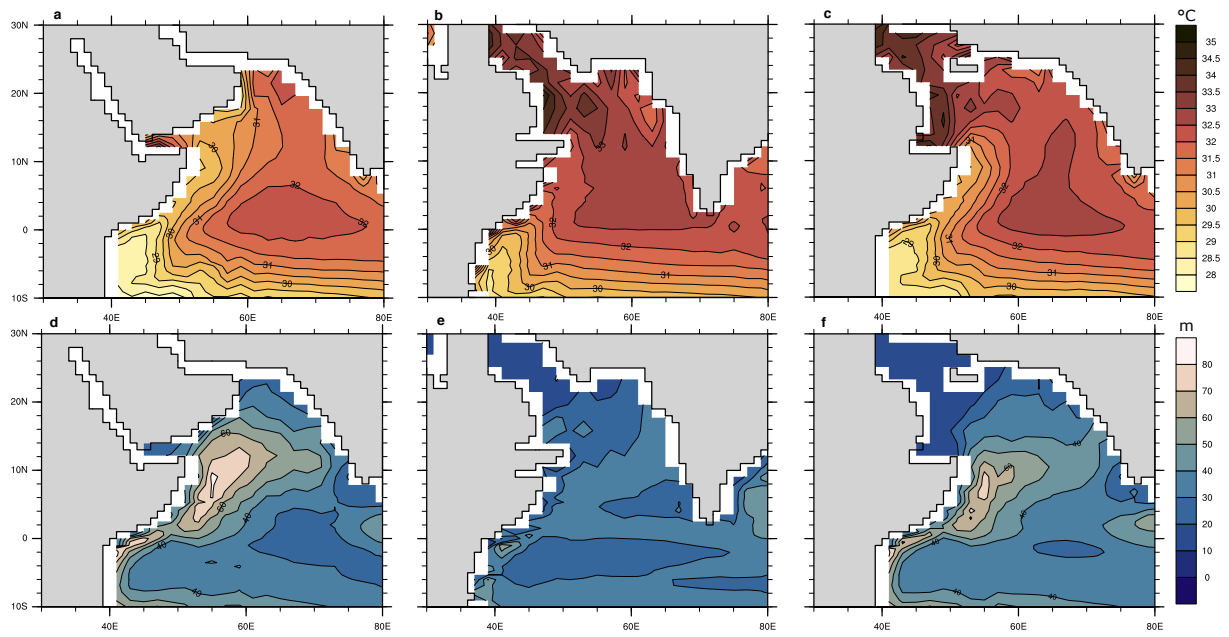


Extended Data Table 1: Simulations performed with IPSL-CM5A2. See Supplementary Figure 1 and 3 for paleogeography maps.

Simulation	pCO ₂ (ppm)	Geography	Sensitivity experiments			
			East Africa	HTP	Anatolia-Iran	East Arabian Plate
EM	560	Early Miocene Paleogeography	Low		Early Miocene	
EM-EAH			From LM			
EM-HTP			Low	From LM		
EM-Him			Low	From LM + 20% higher Himalaya		
LM	560	Late Miocene Paleogeography	100%	Late Miocene	100 %	100 %
LM-NoAIO			100%		50 %	100 %
LM-NoEAP			100%		50 %	0 %
LM-NoEAHR			60%		50%	100%
LM-AIS			100%		100 %	100 %
LM-CO2			420		100%	100 %

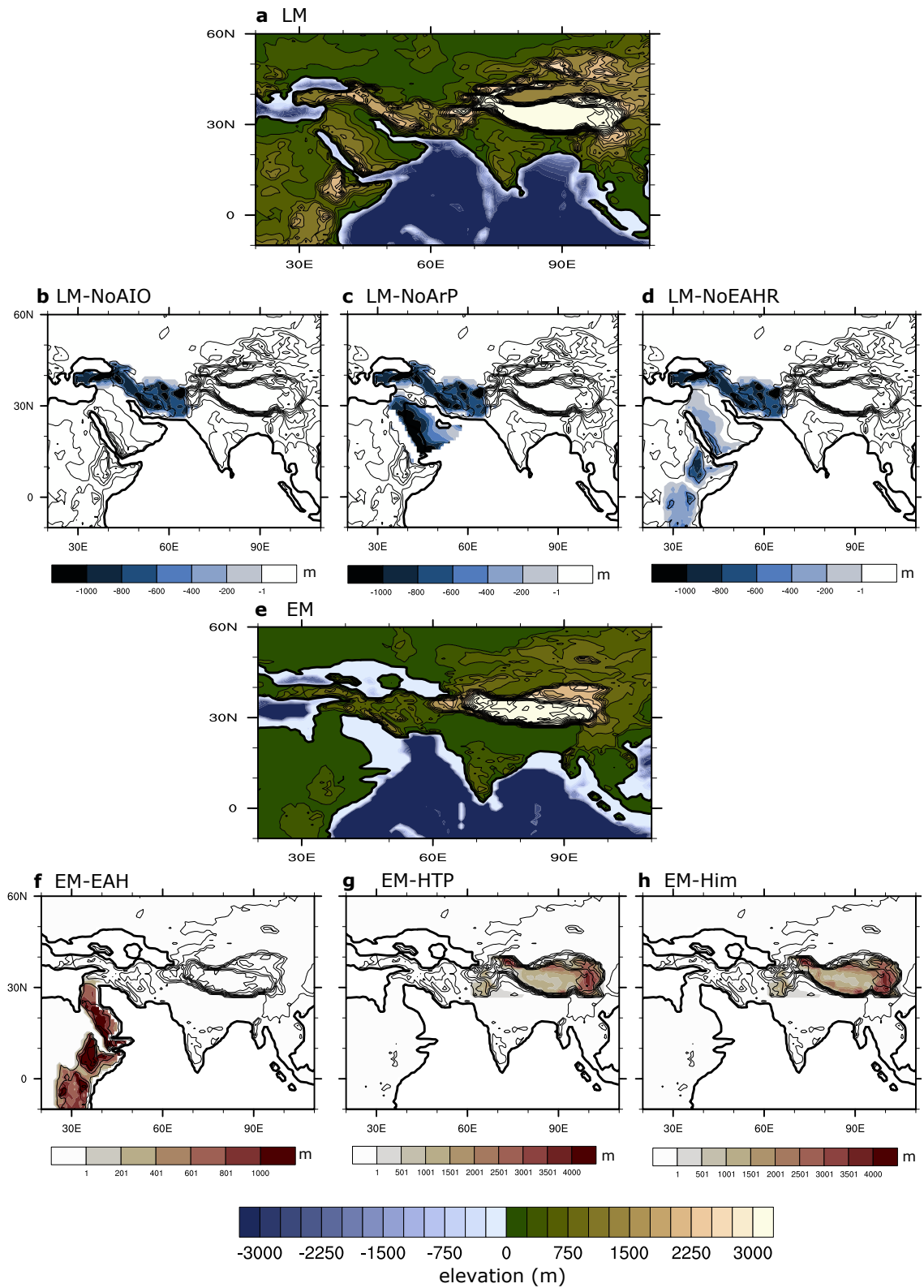


Extended Data Figure 1: Paleogeographic reconstruction used in the reference simulations. a) late Miocene (LM) and b) early Miocene (EM) simulations. Initial bathymetry is more detailed in LM than in EM paleogeography, but the model resolution (2° by 2°) mitigates the difference by smoothing small variations.

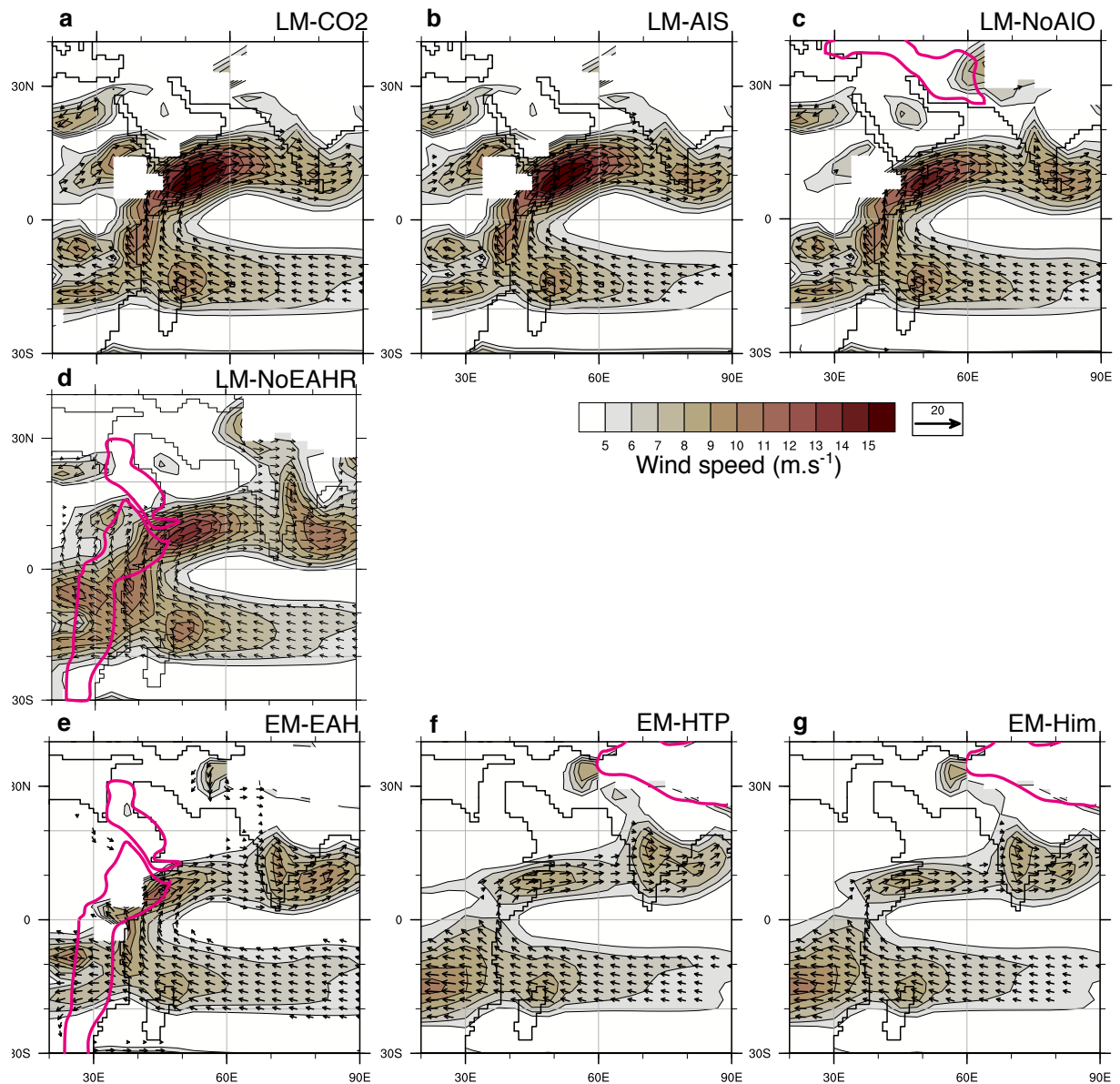


Extended Data Figure 2: Western Indian ocean response to Miocene paleogeographic evolution

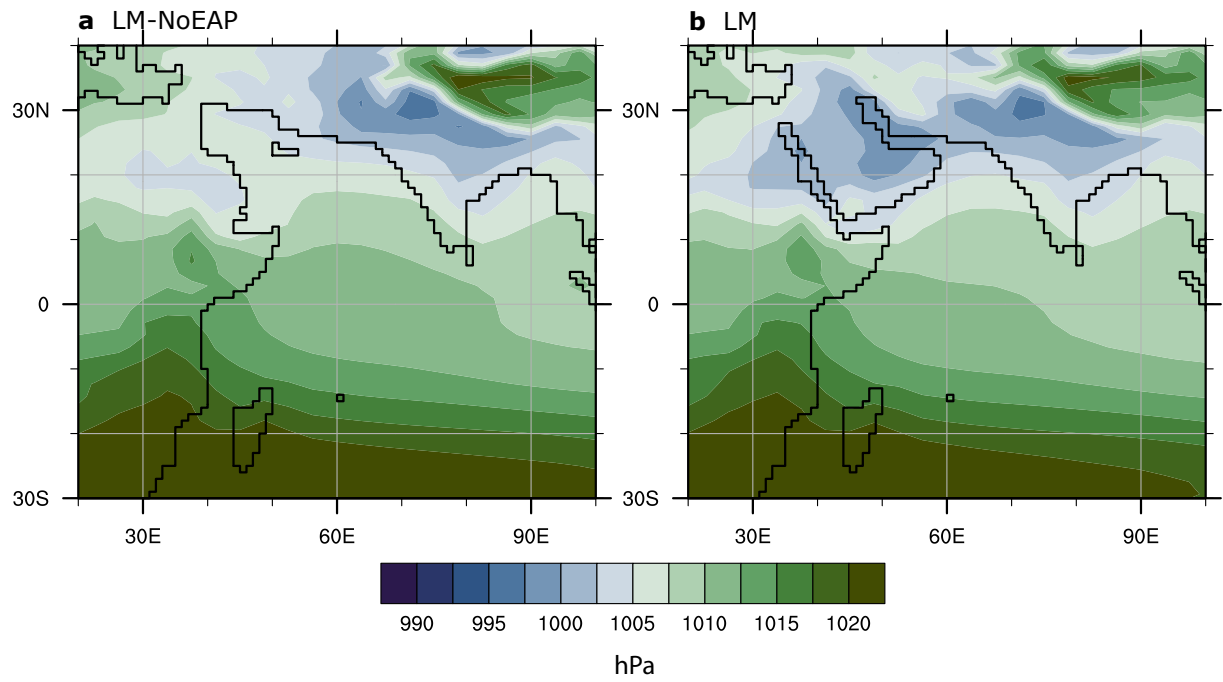
Top : Sea surface temperatures ($^{\circ}\text{C}$) averaged over boreal summer (JAS) ; Bottom : Mixed layer depth average during boreal summer (JAS). (a)(d) late Miocene (LM) and (b)(e) early Miocene (EM) and (c)(f) LM-NoEAP simulations (See Extended Data Tab. 1 and Extended Data Fig. 3). LM-NoEAP is a simulation without the Eastern Arabian Peninsula (EAP), designed to show the influence of Arabian Peninsula immersion on the Somali Jet structure in a LM configuration.



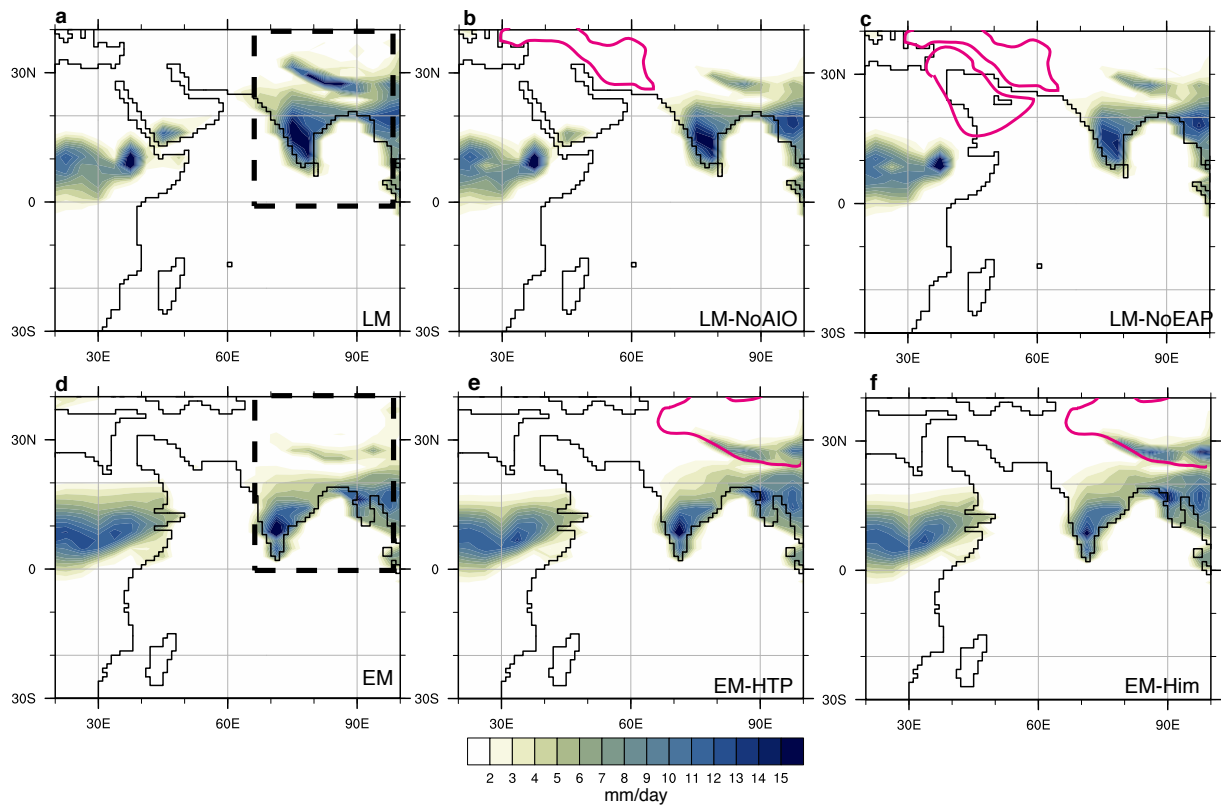
Extended Data Figure 3: Paleogeographic changes used in the sensitivity experiment. Paleogeography surrounding the Indian Ocean in a) the late Miocene (LM) and e) the early Miocene (EM) simulations. Change in paleogeography between sensitivity experiments and the reference simulation (LM and EM) ; b) LM-NoAIO (change in Anatolia-Iran orogen (AIO) topography) ; c) LM-NoEAP (Change in Arabian Peninsula (EAP) land extension) d) LM-NoEAHR (Change in East African Highland (EAH) topography on LM configuration) ; f) EM-EAH (Change in East African Highland topography on EM configuration) ; g) EM-HTP (Change in Himalaya and Tibetan Plateau (HTP) topography) and g) EM-Him (Higher than present-day Himalaya). Contours are drawn every 250 meters for EM and LM simulations and every 500 meters for each sensitivity experiment.



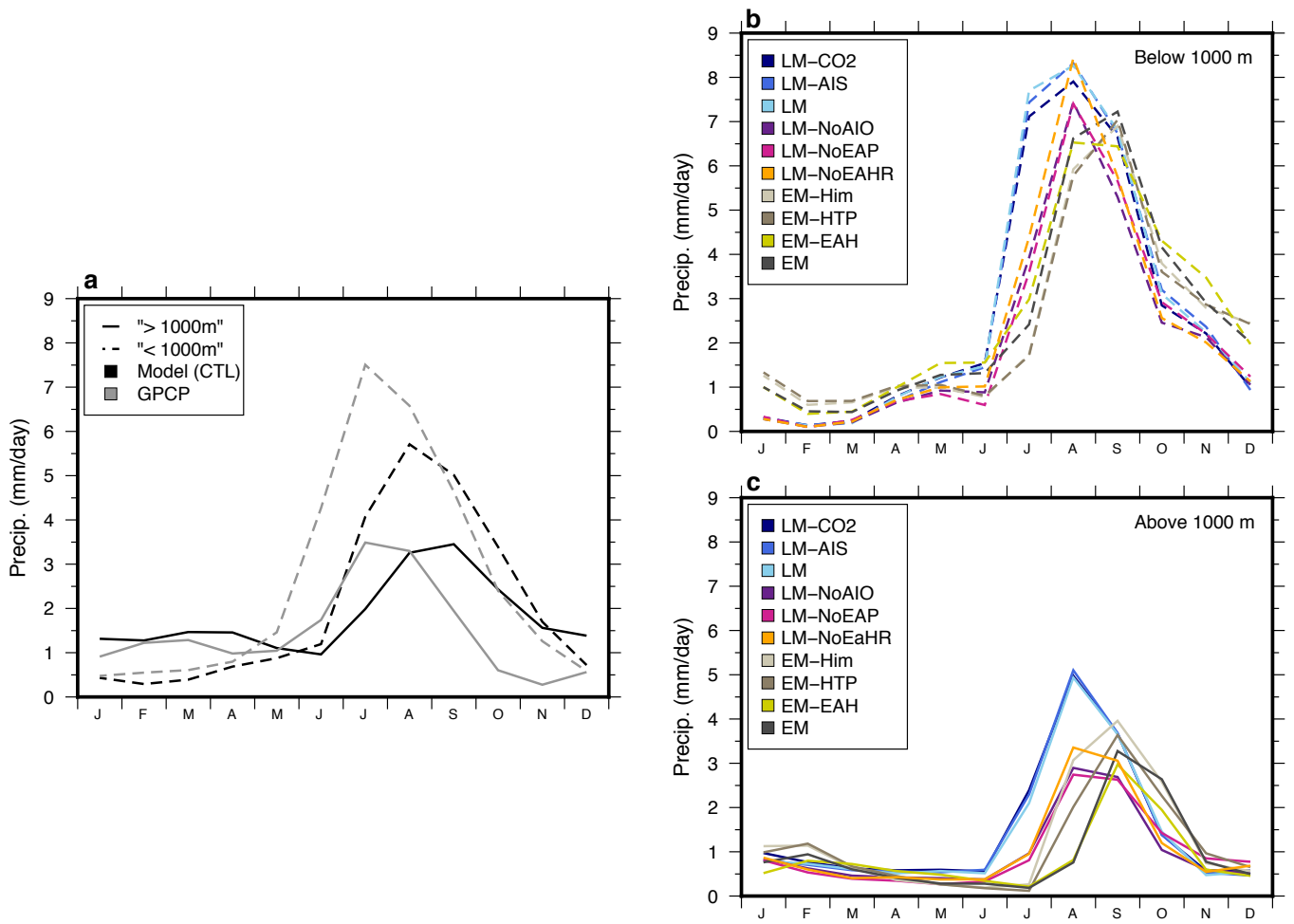
Extended Data Figure 4: Somali Jet response to changes in regional paleogeography. Low level winds (850 hPa) during boreal summer (JJA) for a) late Miocene (LM) with low CO₂ (LM-CO₂); b) LM with Expanded Antarctic Ice Sheet (LM-AIS); c) LM with half present-day Anatolia-Iran topography (LM-NoAIO); d) LM with reduced topography in East African Highlands (LM-noEAHR) ; e) EM with uplifted East African Highlands (EM-EAH); f) EM with fully uplifted HTP region (EM-HTP) and g) EM with higher than present-day Himalaya orography (EM-Him).



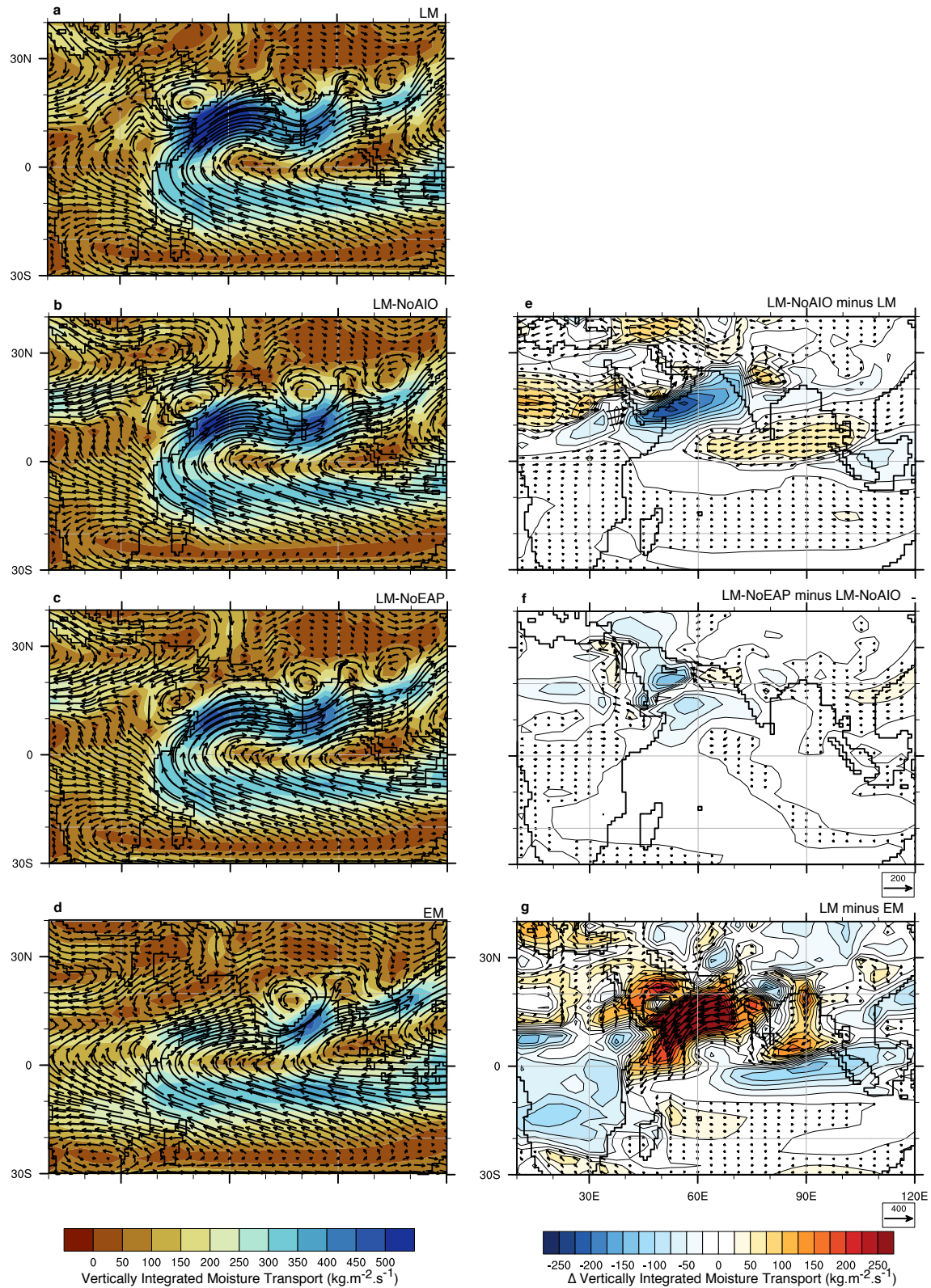
Extended Data Figure 5: Sea level pressure response to Middle Eastern physiographic changes. Mean summer (JJA) sea level pressure (hPa) for a) late Miocene (LM) with partly submerged Eastern Arabian Peninsula (LM-NoEAP) and b) late Miocene baseline simulation.



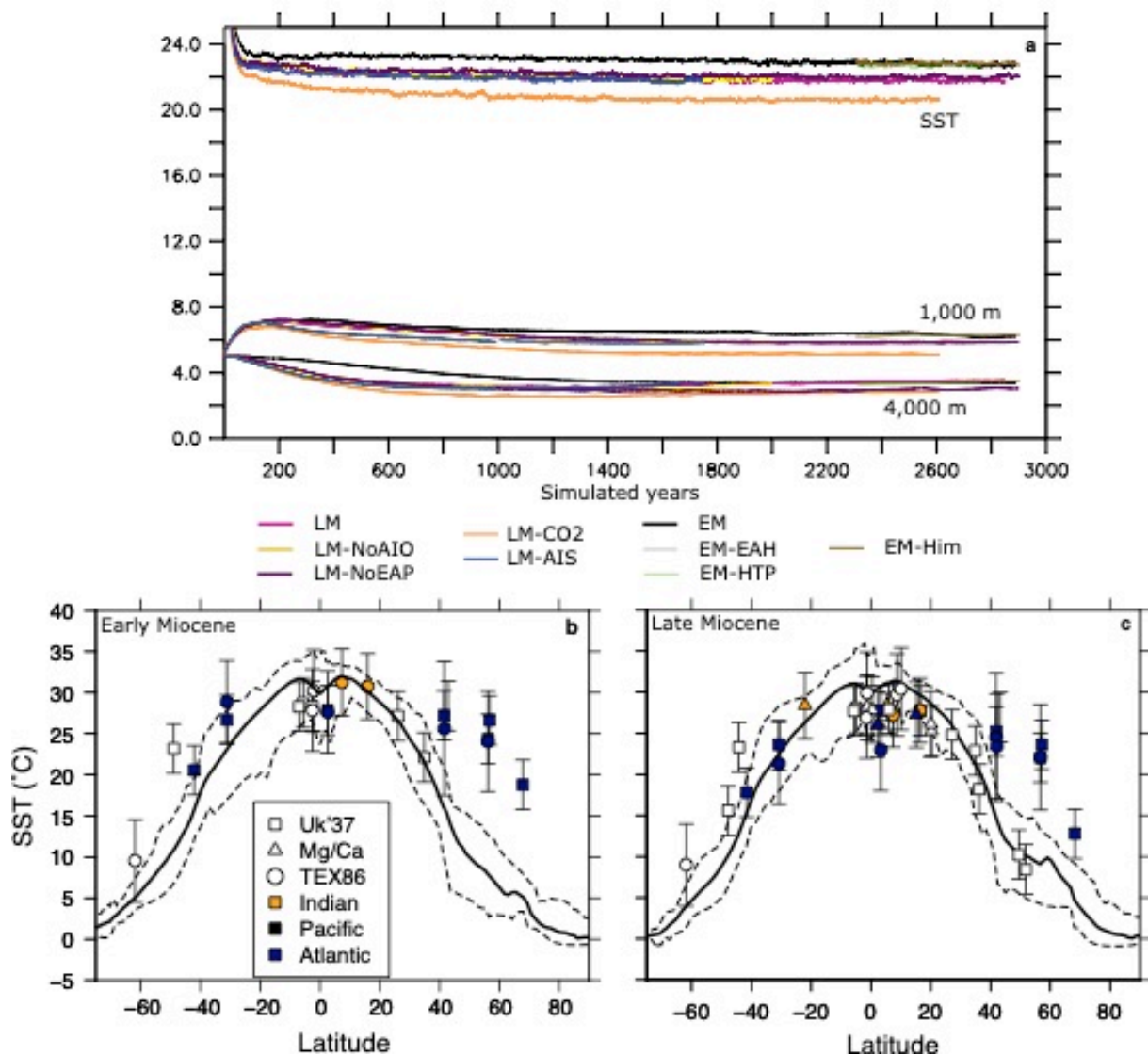
Extended Data Figure 6: Mean summer precipitation response to change in regional physiography. a) late Miocene (LM), b) LM with half present-day Anatolia-Iran topography (LM-AIO), c) LM with partly submerged Eastern Arabian Peninsula (LM-NoEAP) and reduced topography in the Anatolia-Iran region, d) early Miocene, e) EM with fully uplifted HTP region (EM-HTP), f) EM with higher than present-day Himalaya orography (EM-Him). Dashed square indicates the area over which inland precipitation seasonality index (Main text, Figure 4) is computed.



Extended Data Figure 7: Seasonal cycle of precipitation. Precipitation is averaged over [65°E-85°E, 0-35°N]. a) Global Precipitation Climatology Project (GPCP) data⁸² and preindustrial simulation²⁴ (Model), b) Low elevation areas (0-1,000 m) and c) high altitude areas (above 1000m) for early Miocene (EM) and late Miocene (LM) simulations and sensitivity experiments.



Extended Data Figure 8: Moisture transport response to change in regional physiography. Late summer vertically integrated moisture transport (JJA) a) late Miocene (LM); b) LM-NoAIO; c) LM-NoEAP, d) early Miocene (EM). Change in vertically integrated moisture transport (JJA) in response to e) change in Anatolia-Iran topography (LM-NoAIO vs. LM); f) emersion of the Eastern Arabian Peninsula (LM-NoEAP vs. LM-NoAIO) and g) paleogeographic evolution between the early and the late Miocene (LM vs. EM).



Extended Data Figure 9: Simulation equilibrium and zonal mean temperature gradient. a) Global ocean temperature evolution at the sea surface, in intermediate (1,000 m) and deep waters (4,000 m) for late Miocene (LM) and early Miocene (EM) simulations, and sensitivity experiments. Zonal mean SST gradient for EM (b) and LM (c) simulations compared to available proxy estimation. Bold line indicates mean annual SST, dashed line depicts minimum and maximum value for each latitude. Temperature reconstruction are from ref. 45, based on compilation by ref. 83 and additional information on the compilation can be found in ref. 45.