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The role of paleogeography in Asian monsoon evolution: a review and new insights from climate modelling

D. Tardif^{1,2,3,4}, A.-C. Sarr¹, F. Fluteau⁴, A. Licht¹, M. Kaya⁵, J.-B. Ladant³, N. Meijer⁶, Y. Donnadieu¹, G. Dupont-Nivet^{7,8}, C.T. Bolton¹, G. Le Hir⁴, Q. Pillot¹, F. Poblete⁹, P. Sepulchre³, A. Toumoulin¹⁰, W. Banfield¹

¹Aix Marseille Univ, CNRS, IRD, INRAE, CEREGE, Aix-en-Provence, France

²IRD, DIADE, University of Montpellier, Montpellier, France

³Laboratoire des Sciences du Climat et de l'Environnement, LSCE/IPSL, CEA-CNRS-UVSQ, Université Paris-Saclay, 91191 Gif-sur-Yvette, France

⁴Université Paris Cité, Institut de physique du globe de Paris, CNRS, F-75005 Paris, France

⁵Department of Geological Engineering, Middle East Technical University, Ankara, Turkey

⁶Senckenberg Biodiversity and Climate Research Centre (SBiK-F), Senckenberganlage 25, D-60325

Frankfurt am Main, Germany

⁷Géosciences Rennes, UMR CNRS 6118, Univ Rennes, Rennes, France

⁸Institute of Geosciences, Potsdam University, Potsdam, Germany

⁹Departamento de Geología, Facultad de Ciencias Físicas y Matemáticas, Universidad de Chile, Chile

¹⁰Department of Botany and Zoology, Faculty of Science, Masaryk University, Brno, Czech Republic

Key Points:

- Modern Asian monsoonal characteristics appeared sequentially during the Cenozoic, driven by distinct paleogeographic and climatic forcing factors.
- Model results depict monsoon-like precipitation seasonality in parts of Southern and Eastern Asia since the Paleogene.
- Monsoon-like rainfall regime in Paleogene eastern Asia is triggered by the proto-Tibetan Plateau uplift.
- Increasing continentality, due to the retreat of the Paratethys Sea and the emergence of the Arabian Platform above sea level, are instrumental in the establishment of surface temperature gradients that promote moist air advection from the Indian Ocean towards Asia in summer.
- The Anatolian-Iranian and East African landforms contribute to reinforcing the Somali Jet that brings moisture to Southeastern Asia in summer.
- The uplift of the Mongolian Plateau drives the formation of the Siberian High in winter, which broadly reinforces the East Asian winter monsoon and induces aridification of the Gobi region.
- The uplift of the Tian Shan and Pamir mountain ranges blocks the westerly moisture flux, forces the northward migration of the Jet Stream in summer, and enhances inland Asian aridification. Their effect on the South Asian monsoon is likely dependent on the paleoelevation of neighboring landforms, especially the Anatolian-Iranian Plateau.
- Additional constraints on the respective timing of uplift and paleoelevation are key to improving our understanding of monsoon establishment.

Corresponding author: Delphine Tardif, delphine.tardif@lsce.ipsl.fr

Corresponding author: Anta-Clarisse Sarr, anta-clarisse.sarr@univ-grenoble-alpes.fr

Abstract

The Asian monsoons are triggered by complex interactions between the atmosphere, Asian and African orography, and the surrounding oceans, resulting in highly seasonal climate and specific regional features. It was thought that the Asian monsoon was established during the Neogene, but recent evidence for monsoon-like precipitation seasonality occurring as early as the Paleogene greenhouse period challenges this paradigm. The possible occurrence of monsoons in a climatic and paleogeographic context very different from the present-day questions our understanding of the drivers underpinning this atmospheric phenomenon, in particular with regard to its dependence on geography. In this study, we first take advantage of the wealth of new studies to tentatively draw an up-to-date picture of Asian tectonic and paleoenvironmental evolution throughout the Cenozoic. We then analyze a set of 20 paleoclimate simulations spanning the late Eocene to latest Miocene (~ 40 -8 Ma) in order to better understand the evolution of the distinct Asian monsoon subsystems. At odds with the traditional view of a monsoonal evolution driven mainly by Himalayan-Tibetan uplift, our work emphasizes the importance of peripheral mountain ranges in driving the evolution of Asian climate. In particular, the uplift of East African and Anatolian-Iranian mountain ranges, as well as the emergence of the Arabian Peninsula, contribute to shaping the modern South Asian summer monsoon. We also suggest that East Asian monsoon establishment and the aridification of inland Asia are driven by a combination of factors including increasing continentality, the orographic evolution of the Tibetan Plateau, Mongolia, Tian Shan and Pamir, and $p\text{CO}_2$ decrease during the Cenozoic.

1 Introduction

The South and East Asian monsoons (SAM and EAM) are highly seasonal climatic phenomena that today support the livelihoods of billions of people. During summer, the Arabian Platform and the Southern Asian continent overheat compared to adjacent Indian and Western Pacific Oceans. These continental masses become the locus of a wide low pressure belt, leading to a regional amplification of the Inter Tropical Convergence Zone (ITCZ) seasonal latitudinal migration, and to the formation of the Somali Jet along the East African coast (Fig. 1). This strong cross-equatorial flow steers low-level moist air from the Indian Ocean to the Asian continent and triggers nutrient-rich upwelling in the western Arabian Sea (Schott & McCreary, 2001; Lévy et al., 2007; Curry et al., 1992). Over the Bay of Bengal, this large scale circulation splits into two branches (SI Fig. 14 f). First, the Indo-Gangetic Low Level Jet (Acosta & Huber, 2017) bifurcates towards northern India and drives important orographic precipitation on the Himalayan foothills during the South Asian summer monsoon (SASM). The second branch continues towards eastern Asia (SI Fig. 14 f) and ascends over northwesterly air masses related to the Jet Stream in the subtropics (Kong et al., 2017; Molnar et al., 2010; Sampe & Xie, 2010), forming the Meiyu-Bayu front, a band of strong convective precipitation stretching from Myanmar to Japan, characterizing the East Asian summer monsoon (EASM) (Ninomiya & Shibagaki, 2007). The front progressively migrates northward during spring and summer, following the Jet Stream's seasonal displacement relative to the Tibetan Plateau (TP): it is located south of the TP in winter and moves to its northern edge in summer (Schiemann et al., 2009). In winter, the temperature and pressure gradients between the Asian continent and the neighboring oceans reverse, as a consequence of the lower troposphere strong radiative cooling over snow-covered Asia (Jeong et al., 2011; Cohen et al., 2001; Jhun & Lee, 2004). The Siberian High, a wide anticyclone, develops over the Mongolian Plateau (L. Wang & Chen, 2014) (SI Fig. 14 j), flanked by two oceanic low pressure centers, the Aleutian Low (over the Northwestern Pacific Ocean), and the Maritime Continent Low (over the equatorial Western Pacific Ocean). The East Asian winter monsoon (EAWM) results from these pressure patterns, which drive the advection of cold, dry and dust-laden air masses from inland Asia to Eastern and Southern Asia (Fig. 1).

91 The spectacular extent and intensity of the Asian monsoons, combined with the highly
92 dynamic tectonic context of the region, have triggered a lot of interest within the scientific
93 community, in order to understand potential links between orography, land-sea distribution
94 and Asian climate evolution. The Asian monsoons have long been viewed as an atmospheric
95 phenomenon intimately intertwined with the Himalayan-Tibetan Plateau uplift history (e.g.
96 Molnar et al., 1993; Zhisheng et al., 2001; Tada et al., 2016; X. Liu, Guo, et al., 2015).
97 Records of strong upwelling initiating in the Arabian Sea at ~ 8 Ma (Kroon et al., 1991)
98 and, later, evidence for important dust deposition on the Chinese Loess Plateau since ~ 22
99 Ma (Guo et al., 2002) have rooted the origin of the Asian monsoons in the Neogene. However,
100 a wealth of recent studies challenge this paradigm. Aeolian dust deposits in northeastern
101 Tibet, as well as paleobotanical and isotopic data indicating a seasonal climate in Myanmar
102 and China (Licht et al., 2014; Sorrel et al., 2017; Fang et al., 2021; Zheng et al., 2022;
103 Meng et al., 2018; Y. Xie et al., 2022) have been interpreted as evidence for an active
104 monsoon as early as the late Eocene (see Fig.2 for a summary of available paleoclimate
105 indicators). A potential Paleogene origin for the Asian monsoons questions the traditional
106 view of monsoon development in relation to the Himalayan-Tibetan Plateau uplift history.
107 First, because the Paleogene exhibited a generally warmer and ice-free climate compared to
108 the Neogene (Westerhold et al., 2020). And second, because most modern Asian mountain
109 ranges were in the early stages of their uplift history, and Asian continentality was lower
110 due to the presence of the Paratethys Sea to the west (Kaya et al., 2019) and higher global
111 sea level (see late Eocene paleogeography in Fig. 3 a).

112 Climate modeling provides a unique opportunity to individually test the effect of each
113 potential forcing factor on the resulting climate. For instance, model studies have overall
114 tended to highlight that paleogeography out-competes global $p\text{CO}_2$ variations or far-field
115 ice-sheet effects in driving the initiation and intensification of the SAM and EAM over
116 multi-million years time scales (e.g. Farnsworth et al., 2019; J.-Y. Lee et al., 2015; Zoura et
117 al., 2019; Roe et al., 2016; Thomson et al., 2021), independently of the model complexity,
118 resolution, or type of geography (derived from modern or more realistic paleogeography).
119 Sensitivity experiments of increasing complexity and resolution have helped disentangle the
120 effect of the Himalaya-Tibet (and its sub-units) uplift from that of peripheral landforms,
121 such as the Tian Shan ranges, the Anatolian-Iranian orogen, or the Eastern African and
122 Mongolian landforms (H. Tang et al., 2013; Acosta & Huber, 2020; R. Zhang, Jiang, &
123 Zhang, 2017), and from the effect of changes in land-sea distribution (Fluteau et al., 1999;
124 Z. Zhang et al., 2007). While modeling studies provide a mechanistic basis to test po-
125 tential drivers for monsoons and aridity, robust knowledge of the timing and evolution of
126 paleogeography and paleoclimate remains of paramount importance to understanding the
127 complex Cenozoic history of Asian monsoons. The past decade has been rich in both mod-
128 eling and field studies, with for example the extension of marine sediment cores back to the
129 Paleogene in the oceans surrounding Asia (see synthesis by Clift et al. (2022)), the increased
130 coverage of well-dated terrestrial paleoclimate indicators from the Asian continent, and new
131 constraints on Cenozoic tectonic and paleoclimate evolution in Asia.

132 In this contribution, we build upon previous work (Sarr et al., 2022; Tardif et al.,
133 2020; Barbolini et al., 2020) to provide a comprehensive spatial and temporal picture of
134 large-scale wind circulation, precipitation seasonality, and aridity in continental Asia from
135 the late Eocene to the late Miocene, using up-to-date paleoclimate simulations. In the
136 following sections, we first synthesize Asian monsoon history, as understood from proxy and
137 modeling perspectives, and highlight the remaining outstanding questions (section 2, Fig. 2).
138 Then, we compile a summary of the main paleogeographic events that affected continental
139 Asia and the regions surrounding the Indian Ocean during the Cenozoic (section 3). We
140 further introduce the late Eocene to late Miocene paleogeographic configurations used to
141 run the climate model (Fig.3), as well as the criteria used to track Asian monsoon evolution
142 throughout the different simulations (section 4). Our modeling results are analyzed and
143 discussed in light of current knowledge.

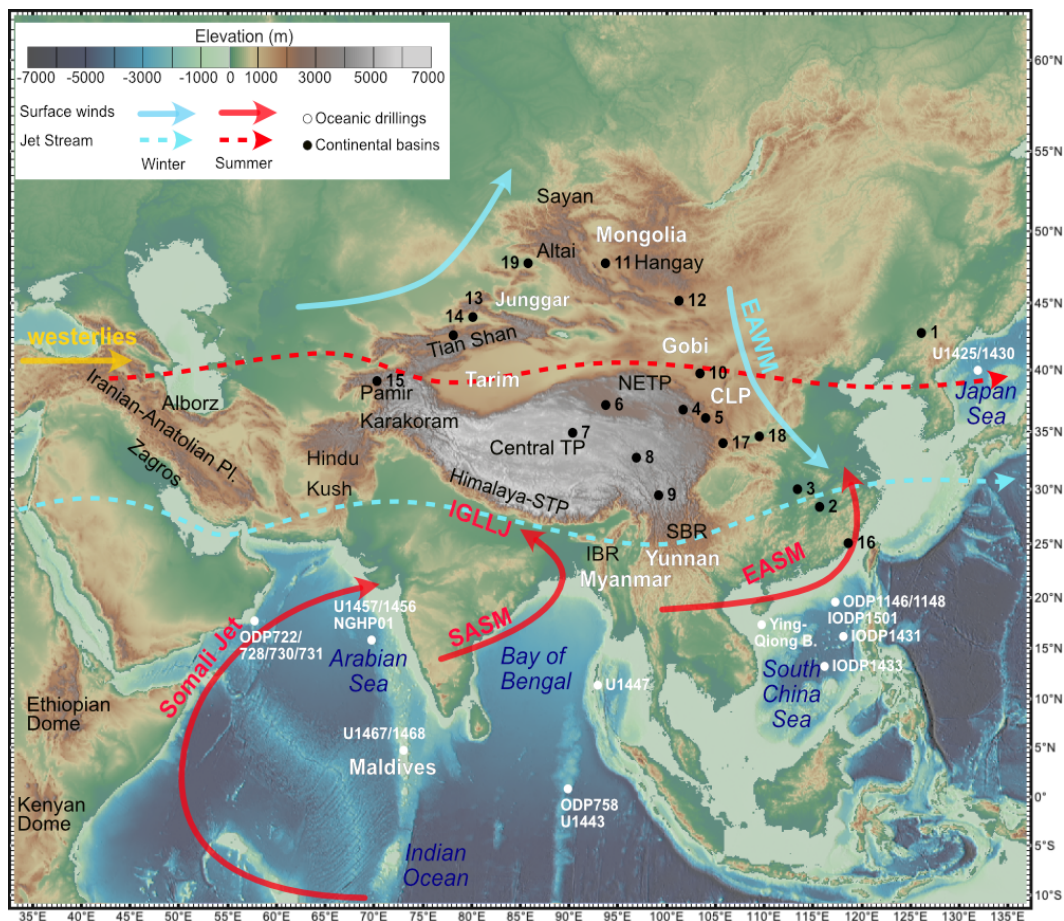


Figure 1. Asian orography (www.geomapapp.org), landforms (black labels), geographical regions (white labels), basins, oceanic drilling sites and seasonal winds, with the following abbreviations: EAWM - East Asian winter monsoon, SASM - South Asian summer monsoon, EASM - East Asian summer monsoon, IGLLJ - Indo-Gangetic Low Level Jet, STP - Southern Tibetan Plateau, NETP - Northeastern Tibetan Plateau, IBR - Indo-Burman Ranges, SBR - Sino-Burman Ranges, CLP - Chinese Loess Plateau. Numbered localities: (1) Huadian B., (2) Qingjiang B., (3) Jiangnan B., (4) Xining B., (5) Lanzhou B., (6) Qaidam B., (7) Hoh Xil B., (8) Nangqian B., (9) Markam B., (10) Bayanhot B., (11) Valley of Lakes, (12) Taatsin Gol, (13) Ili B., (14) Issyk Kul B., (15) Tajik B., (16) Zhangpu, (17) Tianshui B., (18) Weihe B., (19) Zaysan B.

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2 Evidence for increasingly old Asian monsoons

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The classical view of Asian paleoenvironmental evolution during the Cenozoic proposes a transition from a Paleogene "zonal climatic pattern" to a Neogene "monsoonal climatic pattern" (Guo et al., 2008; X. Sun & Wang, 2005; Jia et al., 2003; F. Wu et al., 2022). The zonal pattern is described as a widespread arid to semi-arid band stretching across China and bracketed in its southern and northern borders by two humid belts, whereas the monsoonal pattern consists of a core of pronounced aridity in the Asian interior, surrounded by monsoonal (seasonally wet) climate on its southern and eastern edges. This view is mostly based on paleoclimate proxy compilations (e.g. Boucot et al. (2013)) used to infer either humid (coal, indicators of forested environments, large mammals, etc.) or semi-arid to arid (evaporites, desert and shrub environments, etc.) conditions. Recent evidence for a highly seasonal climate in India, Myanmar and China as early as the Eocene, at least 20

156 Ma earlier than previously documented, nevertheless challenges this paradigm. This section
157 comprehensively reviews the broad paleoenvironmental changes inferred from paleoclimate
158 indicators in Asia from the late Eocene to the late Miocene, which are compiled in Figure 2
159 (upper part). Localities and basins mentioned throughout the text are detailed in Figure 1.

160 **2.1 Eocene to Oligocene paleoenvironmental evolution**

161 *2.1.1 A progressive increase in summer monsoon indicators*

162 The oldest evidence for strongly seasonal rainfall in Southern Asia comes from north-
163 western Indian early Eocene floras (Shukla et al., 2014; Bhatia, Khan, et al., 2021), as well as
164 from mid-late Eocene (~ 40 Ma) floras and isotopic measurements in freshwater gastropod
165 shells and mammal tooth enamel found in Myanmar (Licht et al., 2014; H. Huang et al.,
166 2021). Both regions were however located between $0-10^{\circ}\text{N}$ at this time (Shukla et al., 2014;
167 Westerweel et al., 2019), and such precipitation seasonality is therefore mainly interpreted
168 as a pronounced seasonal ITCZ migration rather than a monsoonal climate. Further east,
169 in the Yunnan region (Southwestern China), which is today subject to a mixture of SAM
170 and EAM influence, a shift from arid/semi-arid to humid environments is recorded during
171 the late Eocene and interpreted as the onset of the Asian monsoon (Sorrel et al., 2017; Fang
172 et al., 2021; Zheng et al., 2022). This climate transition is inferred by the presence of coal
173 layers, a transition from xerophytic to mixed forest pollen assemblage, a change in lithology
174 interpreted as lake and swamp expansions, as well as the occurrence of fossils of freshwater
175 fish and large mammals. A seasonally wet monsoonal climate is also inferred from isotopic
176 measurements in gastropods shells (Fang et al., 2021). The exact timing of this paleoen-
177 vironmental change is still debated between 41 and 35.5 Ma, but it nevertheless predates the
178 Eocene-Oligocene Transition (EOT, ~ 34 Ma). The proposed driving mechanisms of this re-
179 gional climate shift involve either a threshold response to Tibetan Plateau elevation (Zheng
180 et al., 2022), a side effect of the Paratethys Sea retreat in the west (that would have allowed
181 more moisture penetration from the ocean) (Fang et al., 2021), or the global climate cooling
182 initiated in the Middle Eocene and subsequent sea surface temperature changes (Sorrel et
183 al., 2017).

184 Evidence for the presence of an EASM during the Eocene is also debated. In east-
185 ern China, paleovegetation mostly points to warm, humid and weakly seasonal evergreen
186 forested environments close to the coast (Spicer et al., 2016; X. Ma et al., 2012; Pound
187 & Salzmann, 2017), except for late Eocene floras from Huadian (Meng et al., 2018) and
188 Qingjiang Basins (Y. Xie et al., 2022), situated in northeastern and eastern China respec-
189 tively, that already indicate highly seasonal rainfall. Further inland, towards the eastern
190 TP, a gradual transition from humid to subhumid and/or seasonally dry paleoenvironments
191 is suggested by botanical evidence (Han et al., 2022; Q. Li et al., 2022) and sedimentary
192 records from various basins displaying an alternation of mudstone and evaporite deposits
193 (D. Wang et al., 2013; Abels et al., 2011). High resolution deposits from the Jiangnan Basin
194 spanning the mid-Eocene to early Oligocene ($\sim 40-34$ Ma) (C. Huang & Hinnov, 2019) show
195 that the penetration of moisture into eastern China in the late Eocene was likely strongly
196 modulated by orbital forcing. These pieces of evidence suggest that the classical Paleo-
197 gene wet/dry zonal climatic pattern (Guo et al., 2008; X. Sun & Wang, 2005) was already
198 periodically disturbed in the late Eocene, preceding the more widespread appearance of
199 monsoon-like precipitation seasonality in the Oligocene.

200 Evidence for increased rainfall seasonality during the Oligocene to early Miocene be-
201 comes more common in floras from both China (X. Ma et al., 2012; Herman et al., 2017;
202 Vornlocher et al., 2021; J. Ren et al., 2021; Miao et al., 2013; Ling et al., 2021; S. Li et al.,
203 2018; C. Huang & Hinnov, 2019; H. Tang et al., 2020) and India (Bhatia, Khan, et al., 2021;
204 Spicer et al., 2017; Srivastava et al., 2012). Additionally, paleoceanographic records from
205 the adjacent Bay of Bengal and South China Sea have been used to track the SAM and
206 EAM evolution from the middle Oligocene onwards. Records from the Ying-Qiong Basin

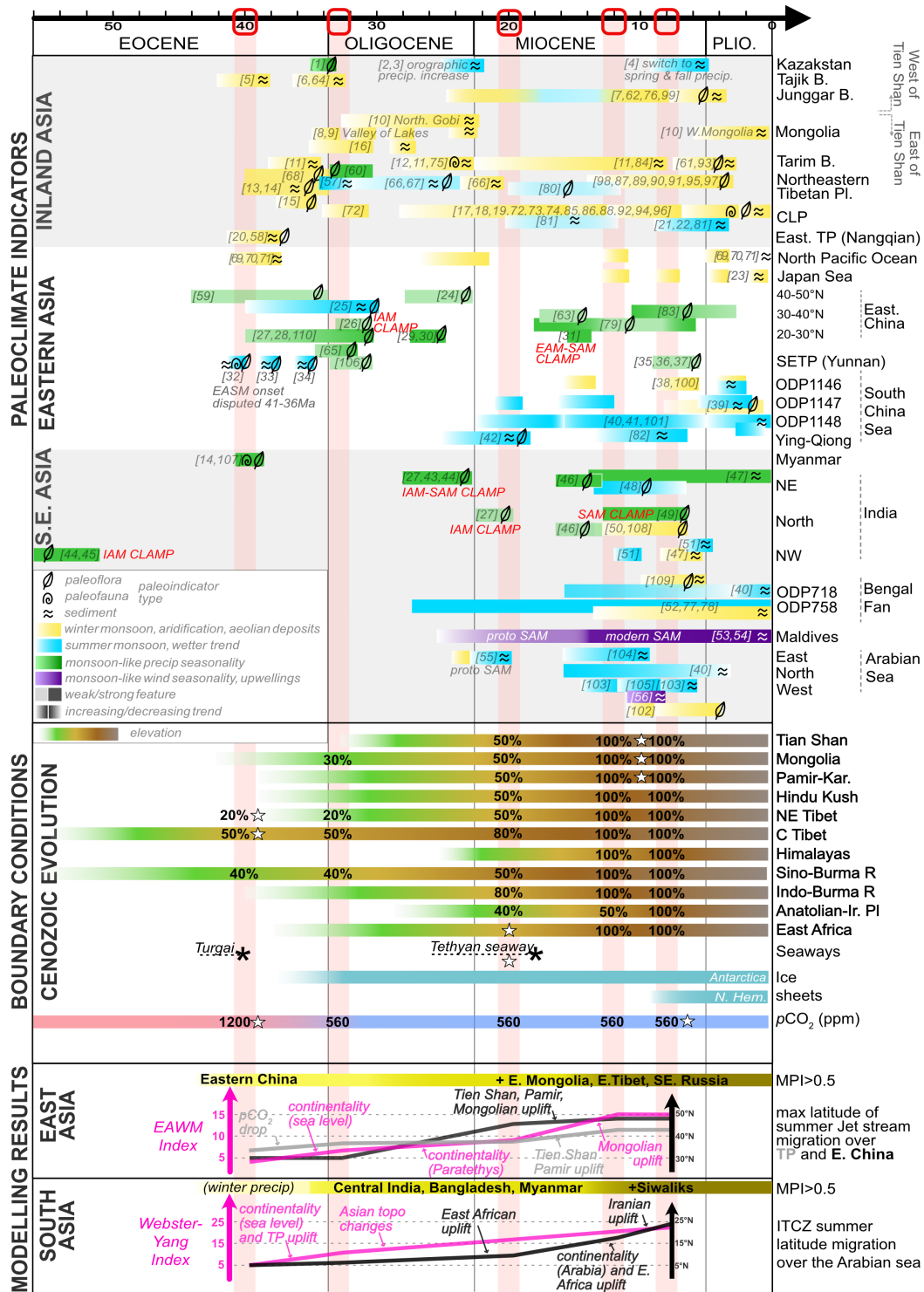


Figure 2. upper Paleoclimate trends recorded in Asia. Bibliographic references in brackets are listed in SI Table 3 ; center Evolution of main landforms, seaways, ice sheets and pCO₂ during the Cenozoic overlain with values used in our reference simulations (elevations expressed in % of modern). White stars indicate parameters that were tested in the sensitivity experiments discussed in the text; bottom Overview of the main monsoon indicator evolution computed from simulations presented in this study. We use the following abbreviations: NETP (SETP) - Northeastern (Southeastern) Tibetan Plateau, CLP - Chinese Loess Plateau.

(South China Sea) show an increase in terrigenous organic matter influx together with a higher proportion of tropical-subtropical angiosperm pollen at ~ 25 Ma, interpreted as a sign of increasing EASM intensity (W. Ding et al., 2021). In the Bay of Bengal (ODP Site 758), the radiogenic isotopic composition of clays indicates relatively constant weathering patterns over the last 27 Ma, and by inference stable SASM activity since at least the late Oligocene (S. Ali et al., 2021).

2.1.2 *Late Eocene aridification in response to the Paratethys Sea retreat and global climate cooling*

Markers of subhumid to arid climate are widespread in inland Asia in the Eocene, with many basins (e.g. Xining, Hoh Xil, Qaidam and Tarim) recording pollen assemblages characteristic of a steppe-desert environment (Q. Yuan et al., 2020; Miao, Wu, et al., 2016; X. Ma et al., 2012). During the mid-late Eocene (~ 40 -34 Ma), inland Asia evolves towards even drier conditions, as suggested by increased proportions of xerophytic plants such as *Nitraria* and *Ephedra* in eastern and northeastern Tibet and in the Tarim Basin (Q. Yuan et al., 2020; Barbolini et al., 2020; Hoorn et al., 2012; Miao, Wu, et al., 2016), $\delta^{18}\text{O}$ measurements on oyster shells in the Tarim Basin (Bougeois et al., 2018), decreasing saline lakes in the Xining Basin (Abels et al., 2011), reduced chemical weathering observed in the Qaidam and Xining Basins (Fang et al., 2019), $\delta^2\text{H}$ isotope analyses on sedimentary leaf waxes *n-alkanes* in the Qaidam Basin (M. Wu et al., 2021) and lipid biomarkers and carbon isotopic compositions from bulk sediments in the Nangqian Basin (J. Wei et al., 2022). The presence of detrital material identified as Asian dust in sediments from the central North Pacific (GPC3 and ODP Site 1215) since ~ 40 Ma also likely indicates the presence of dry conditions on land (Pettke et al., 2002; D. Rea et al., 1985; Ziegler et al., 2007).

This late Eocene aridification episode could result from a reduced westerly moisture input following two successive Paratethys Sea regression phases from the Tarim and the Tajik Basins between ~ 41 -37 Ma (Carrapa et al., 2015; J. Sun et al., 2020, 2022; Bougeois et al., 2018; Bosboom et al., 2014b; Kaya et al., 2019). High resolution sedimentary records, such as those of the Xining Basin (northeastern Tibet), exhibiting good correlation between wetter saline lake deposits during Paratethys Sea transgressions (Meijer et al., 2019; Bosboom et al., 2014a), confirm the key role played by the westerlies and the Paratethys Sea as a major moisture source for inland Asia in the Eocene. Additionally, the strong obliquity cyclicity imprinted on Xining evaporite/mudflat alternations in the late Eocene (40-34 Ma), coeval with the appearance of loess-like dusts in the alluvial mudflats (Meijer et al., 2021; Licht et al., 2014), hint at a marked influence of high latitude dynamics (incipient polar ice-sheets, variations in the Siberian High intensity) on winter inland Asian climate at that time (Abels et al., 2011; Xiao et al., 2010; Meijer et al., 2021).

2.1.3 *Oligocene climate evolution and inception of modern Asian deserts*

In the early Oligocene, a few sites situated within the westerly moisture area display increased humidity, recorded by $\delta^{18}\text{O}$ isotopic measurements in pedogenic carbonates in the Tarim and western Qaidam Basins (Kent-Corson et al., 2009), together with $\delta^2\text{H}$ measurements in leaf waxes (M. Wu et al., 2021). Leaf fossil CLAMP analysis (Climate Leaf Analysis Multivariate Program (Feng & Poulsen, 2016; Wolfe, 1993; Jacques, Su, et al., 2011)), further suggest a rather wet (but not monsoonal) rainfall regime in western Qaidam (Song et al., 2020). On the other hand, evidence for a drier climate is reported within the EAM domain by $\delta^{18}\text{O}$ isotopic measurements in pedogenic carbonates from eastern Qaidam (Y. Sun et al., 2020; Kent-Corson et al., 2009) and in Lanzhou Basin in the Chinese Loess plateau area (B. Li et al., 2016). CLAMP analysis on fossil leaves in the Lanzhou (Miao et al., 2013) and southeastern Tibetan Markam Basin (T. Su, Spicer, et al., 2019) advocate for an increased (and monsoon-like) precipitation seasonality due to the TP uplift. Interpretations for this climatic trend in eastern China suggest that it may indicate either decreased EASM rainfall due to global cooling after the EOT (M. Wu et al., 2021) or increased winter

258 aridity due to the Paratethys Sea retreat and/or to global sea level fall after the EOT (B. Li
259 et al., 2016).

260 In Inner Mongolia (Taatsin Gol region), $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ isotopic measurements in
261 pedogenic carbonates indicate aridification since the early Oligocene (Caves Rugenstein et
262 al., 2014) and possible loess occurrence as early as ~ 34 Ma (J. Sun & Windley, 2015).
263 Proposed mechanisms for this drying involve increased continentality due to Paratethys
264 Sea retreat at 34 Ma (J. Sun & Windley, 2015) and an early Hangay Dome uplift that
265 would have acted as a barrier to moisture advection from Siberia to Mongolia as early as
266 the Oligocene (Caves Rugenstein et al., 2014). However, high resolution records reveal
267 several pulses of aridification at ~ 34 -33, ~ 31 , ~ 28 and ~ 23 Ma, coeval with periods of
268 ice-sheet expansion, suggesting that these cooling phases were likely the main drivers of
269 Mongolian hydrological dynamics (Baldermann et al., 2021). Further south, in the Gobi
270 region (Bayanhot Basin), two bursts of aridification inferred from environmental magnetic,
271 mineralogical and geochemical data at ~ 31 and ~ 28 Ma further hint at an influence of high
272 latitude ice-sheet fluctuations on Mongolian Oligocene climate (Wasiljeff et al., 2022).

273 In the mid to late Oligocene, eolian dust deposits are documented in three major
274 sites (see Meijer et al. (2021) for a review), namely the Chinese Loess Plateau, from ~ 29
275 Ma onwards (Garzzone et al., 2005; Qiang et al., 2011), the Tarim Basin at ~ 27 -22 Ma
276 (Zheng et al., 2015), and the Junggar Basin at ~ 24 Ma (J. Sun et al., 2010). Aridification
277 is also inferred from $\delta^{18}\text{O}$ isotopic measurements in oysters and pedogenic carbonates from
278 the Tarim Basin at ~ 25 Ma (Bougeois et al., 2018; Kent-Corson et al., 2009) and from
279 isotopic records in stromatolites from the Junggar Basin, loosely dated to the early Miocene
280 (W. Yang et al., 2019). This second step of aridification is coeval with $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ isotopic
281 measurements from pedogenic carbonates suggesting increased orographic precipitation in
282 the Issyk Kul and Ili Basins, both situated on the western side of the Tian Shan ranges
283 (Hellwig et al., 2018; Macaulay et al., 2016). Collectively, these observations are interpreted
284 as an indication of an uplift episode of the Tian Shan-Pamir ranges, which would have
285 started shielding the Tarim and Junggar Basins from westerly moisture input and favored
286 their aridification (Bougeois et al., 2018; W. Yang et al., 2019). This late Oligocene to
287 early Miocene aridification of the Asian interior is also depicted in several oceanic records,
288 where an increase in dust mass accumulation rate is recorded in the Pacific at ~ 25 Ma (Site
289 GPC3) (D. K. Rea, 1994; Pettke et al., 2002) and at ~ 22 Ma (ODP Site 1215) (Ziegler et
290 al., 2007). A reduction in $\Delta\delta^{13}\text{C}$ coeval to an increase in bulk sediment Mn/Fe and Mn/Ti
291 and foraminiferal Mn/Ca in the eastern Arabian Sea at ~ 23 Ma (NGHP01 Site), reflecting
292 an increase in water column mixing and ventilation, are considered as further evidence for
293 the existence of a proto winter monsoon wind system (Beasley et al., 2021).

294 **2.2 Evolution of seasonal and regional monsoons in the Neogene**

295 The better coverage in both continental and marine records in the Neogene allows
296 drawing a more precise picture of monsoon evolution during the Late Cenozoic. This section
297 reviews the many pieces of evidence which now tend to support a scenario of diachronous
298 evolution of regional monsoons, with a decoupling between summer and winter seasons as
299 well as between monsoon-like precipitation seasonality and monsoon-like wind circulation
300 establishment.

301 **2.2.1 Neogene South Asian monsoon variations**

302 The occurrence of monsoon-like rainfall seasonality in the eastern SAM domain is at-
303 tested to by relatively constant isotopic composition of clays derived from silicate weathering
304 since ~ 27 Ma in the central Bay of Bengal (ODP Site 758) (S. Ali et al., 2021). Middle
305 Miocene (~ 13 Ma) plant macrofossils from northeasternmost India (Arunachal Pradesh)
306 also attest to modern-like rainfall seasonality at that time (Khan et al., 2014). In Yunnan,
307 the disappearance from the fossil record of *Metasequoia*, a plant whose modern descendant

is intolerant to dry winters, may further attest to a SAM rainfall seasonality intensification after the middle Miocene (L. Wang et al., 2019). On the other hand, plant macrofossils and pollen from the northwestern SAM domain (Nepal, northwestern India and Pakistan) show a progressive vegetation shift from evergreen to deciduous forests from the middle to the latest Miocene, and finally to C4 grasslands (Hoorn et al., 2000; Bhatia et al., 2022; Bhatia, Srivastava, et al., 2021; Srivastava et al., 2018; Tauxe & Feakins, 2020; Vögeli et al., 2017; Sanyal et al., 2004). This vegetation change is also recorded in bulk sediments, leaf waxes, pollen and macrofossils in the Arabian Sea at \sim 8-7 Ma (ODP Site 722 and IODP Site U1457, (Feakins et al., 2020; Y. Huang et al., 2007; Clift et al., 2022)), and in the Bengal fan at \sim 7-6 Ma (ODP Sites 717/718, (Polissar et al., 2021)).

Chemical weathering and sediment accumulation rates from both the Indus Fan and the Bay of Bengal suggest that SASM precipitation reached a maximum in the middle Miocene (\sim 16-10 Ma) before declining between \sim 8-3 Ma (Clift et al., 2008, 2019; Zhou et al., 2021). This has been suggested to reflect an intensification of the SASM (S. Ali et al., 2021), possibly concomitant with a weakening of the SASM (J. Lee et al., 2020). Wind proxies also point to the presence of a seasonal wind reversal, and corroborate the existence of a winter monsoon since the middle Miocene. In the Arabian Sea, an increased relative abundance of the foraminifera *Globigerina Bulloides* in the sediments was initially used to infer an intensification (or an onset) of the SASM wind circulation at \sim 8 Ma (Kroon et al., 1991), based on the assumption that stronger monsoonal winds would translate into stronger upwelling activity. Those records have since been extended back in time (Bialik et al., 2020; Zhuang et al., 2017; Gupta et al., 2015) and combined with the analysis of current-controlled drift sediments and geochemical tracers from the Maldives archipelago (Betzler et al., 2016). These new studies now point to the establishment of a proto SAM wind circulation in the late Oligocene (\sim 25 Ma), and of its reinforcement in the middle Miocene (\sim 13-10 Ma) (Betzler et al., 2016; Zhuang et al., 2017; Bialik et al., 2020). Strong SAM wind circulation since at least 10 Ma is also supported by export productivity and sediment accumulation patterns from records in the southern Bay of Bengal (IODP Site U1443) (Bolton et al., 2022), together with clay mineralogy in the Andaman Sea (IODP Site 1447) (J. Lee et al., 2020) and isotopic and chemical analyses on sediments in the eastern Arabian Sea (IODP Site 1456) (Tripathi et al., 2017)). An intensification of the SAM in the late Miocene (\sim 8-7 Ma) has also been inferred based on an increase in percent in *G. Bulloides* and Total Organic Carbon content in sediments from Oman margin (Gupta et al., 2015) but remains debated, as it is not observed in all paleoceanographic records from the Arabian Sea and the Bay of Bengal (Bolton et al., 2022; Tripathi et al., 2017; Betzler et al., 2016; Y. Huang et al., 2007).

2.2.2 East Asian Summer Monsoon evolution in the Neogene

Summer precipitation typical of the EASM likely occurred throughout the Neogene, as shown by records from the South China Sea such as weathering indices (Clift et al., 2008, 2014), $\delta^{13}\text{C}$ measurements in black carbon (Jia et al., 2003), pollen, biomarkers, kerogen maceral composition and clay mineralogy (W. Ding et al., 2021). On land, $\delta^{18}\text{O}$ values from paleosol carbonates in the Chinese Loess Plateau (Suarez et al., 2011) and most Chinese paleobotanical and palynological data suggest summer precipitation amounts comparable to modern (Q. Wang et al., 2021; Yao et al., 2011; Xing et al., 2012; Hui et al., 2021; B. Wang et al., 2021; Q.-g. Sun et al., 2002). Many localities nevertheless point to wetter winters due to weaker winter monsoons until the late Miocene (Q. Wang et al., 2021; Liang et al., 2022), except for the mid-Miocene Zhangpu biota flora (southeastern China), whose CLAMP signature shows affinities with EAM and SAM floras, thus suggesting adaption to modern-like rainfall seasonality (B. Wang et al., 2021). A warm and wet period has been identified during the Middle Miocene Climatic Optimum (\sim 17-14 Ma) in both continental as well as oceanic records and was likely less seasonal than during the rest of the Miocene (Clift et al., 2022). The penetration of EASM moisture would have reached as far as the Qaidam Basin, as suggested by lotus fossils (Luo et al., 2022), increased abundance of *Fupingopollenites*, an

361 extinct palynomorph highly dependent on summer moisture (Miao, Song, et al., 2016), and
 362 characterization of iron oxides from the Chinese Loess Plateau (H. Zhao et al., 2020). This
 363 warm and wet mid-Miocene climate peak is also inferred from chemical weathering proxies
 364 and indicators of stronger physical erosion and fluvial input measured in sediments from the
 365 South China Sea at 15-12 Ma (Clift et al., 2014), and from $\delta^{13}\text{C}$ in black carbon at 15 Ma
 366 (Jia et al., 2003).

367 The variations in EASM intensity during the late Miocene and Pliocene remain con-
 368 troversial in both continental and marine records, with variations in the suggested timing of
 369 changes (Clift et al., 2022). A decline in EASM intensity induced by decreasing $p\text{CO}_2$ after
 370 the Mid-Miocene Climate Optimum is suggested by weathering, paleoceanographic indices
 371 and pollen from the South China Sea (ODP Site 1146) (Clift et al., 2014; Holbourn et al.,
 372 2021, 2018; Miao et al., 2017; G. Wei et al., 2006), as well as pollen records from the Liupan
 373 Mountains (North China) (Jiang & Ding, 2008). On the other hand, in the South China
 374 Sea, late Miocene EASM intensification is inferred from $\delta^{18}\text{O}$ and Mg/Ca derived records
 375 of surface and sub-surface water characteristics (IODP Site U1501) (C. Yang et al., 2021)
 376 and by black carbon $\delta^{13}\text{C}$ (Jia et al., 2003). In China, wetter conditions during the lat-
 377 est Miocene are suggested by mineralogical, chemical and magnetic analysis of sedimentary
 378 deposits in the Chinese Loess Plateau (Ao et al., 2016, 2021; H. Zhao et al., 2020), pollen
 379 records from the Tianshui (Hui et al., 2021) and the Weihe Basins (L. Zhao et al., 2020),
 380 and by fossil wood, mammals and fish in the Qaidam Basin (Cheng & Yang, 2016; X. Wang
 381 et al., 2007). EASM intensification at this time is usually attributed to global warming (Ao
 382 et al., 2021; H. Wang et al., 2019) and/or to Neogene TP uplift phases (Hui et al., 2021;
 383 X. Ren et al., 2020; H. Zhao et al., 2020).

384 ***2.2.3 Miocene-Pliocene aridification pulses in response to regional uplift***

385 Throughout the Miocene, dust deposition continued in the Chinese Loess Plateau
 386 (Guo et al., 2002; Meijer et al., 2021), Tarim-Taklimakan (Kent-Corson et al., 2009; Zheng
 387 et al., 2015; Heermance et al., 2018) and Junggar Basins (J. Sun et al., 2010). After a
 388 phase of relatively wetter conditions during the Mid-Miocene Climate Optimum (see section
 389 2.2.2), a pronounced step-wise aridification was then recorded in most of inland Asia until
 390 the Miocene-Pliocene boundary (Lu et al., 2010; Lu & Guo, 2014; Z.-H. Tang & Ding,
 391 2013). The timing of aridification pulses varies depending on the region and is attributed
 392 to either northeastern TP uplift (Miao et al., 2012), Altai and Tian Shan mountain uplift
 393 (Caves Rugenstein et al., 2014, 2017) and/or to late Miocene cooling (Lu et al., 2010; Lu
 394 & Guo, 2014; Peng et al., 2016). In the eastern part of inland Asia, encompassing the
 395 Chinese Loess Plateau, central China and northeastern TP, data provide strong evidence
 396 for aridification, including pollen and biomarkers documenting a step-wise transition from
 397 a sub-humid to an arid environment starting at ~ 15 Ma (J. Liu et al., 2016; Miao et al.,
 398 2011; Jiang & Ding, 2008; Peng et al., 2016), massive grassland expansions at $\sim 11-7$ Ma
 399 (Suarez et al., 2011; Barbolini et al., 2020; Y. Ma et al., 2005; L. Zhao et al., 2020; H. Wang
 400 et al., 2019), widespread eolian dust deposition at $\sim 8-7$ Ma (X. Ma & Jiang, 2015; Guo
 401 et al., 2002; Qiang et al., 2011; B. Li et al., 2016; Jiang et al., 2017), increased $\delta^{18}\text{O}$ of
 402 pedogenic carbonates and terrestrial mammal tooth enamel at $\sim 14-13$ and $\sim 7-5$ Ma (B. Li
 403 et al., 2016; Kent-Corson et al., 2009; M. Wu et al., 2021; W. Liu et al., 2014; Y. Sun et
 404 al., 2020; Kaakinen et al., 2006; Y. Wang & Deng, 2005), n -alkane analysis from Qaidam
 405 sediments since 13 Ma (Z. Liu et al., 2014) and the dominance of cold-aridiphilous mollusks
 406 in the Chinese Loess Plateau from 7-5 Ma (F. Li et al., 2008). The Tarim and Junggar
 407 Basins also provide multiple evidence for mid to late Miocene aridification, such as increased
 408 proportions of xerophytic plants and grass pollen at $\sim 7-5$ Ma (J. Sun et al., 2010; J. Sun
 409 & Zhang, 2008; J. Sun et al., 2008; Z. Zhang & Sun, 2011; Barbolini et al., 2020), massive
 410 aeolian deposition in the western Tarim Basin at $\sim 12-7$ Ma (Heermance et al., 2018) and the
 411 permanent drying of lakes in the eastern Tarim Basin at ~ 5 Ma (W. Liu et al., 2014). The
 412 pedogenic carbonate $\delta^{18}\text{O}$ records of these regions nevertheless offer diverging patterns of
 413 evolution, that have tentatively been explained by their locations with respect to the Tian

414 Shan and Altai mountains (Caves Rugenstein et al., 2017). Records from regions today lying
 415 upwind of the Tian Shan and Altai mountains, such as the Issyk Kul (Macaulay et al., 2016),
 416 Junggar (Charreau et al., 2012) and Zaysan Basins (Caves Rugenstein et al., 2017) exhibit
 417 a decline in $\delta^{18}\text{O}$ between 10-5 Ma, which has been interpreted as a shift to spring/fall
 418 precipitation following the uplift of the Altai and Tian Shan mountains (Caves Rugenstein
 419 et al., 2017). Perennial lake formation in the windward Ili Basin (Kazakhstan) at ~ 10 Ma
 420 may further attest that the Tian Shan ranges acted as an orographic barrier for westerlies
 421 moisture input (Frisch et al., 2019). On the other hand, records from the Tarim Basin,
 422 situated downwind of the Tian Shan ranges, display $\delta^{18}\text{O}$ trends similar to those of the
 423 Qaidam Basin (Kent-Corson et al., 2009) with increased values since ~ 15 Ma peaking at
 424 ~ 7 -5 Ma and which are interpreted as indicating aridification at that time.

425 EAWM intensification is also recorded in the South China Sea at ~ 15 -12, ~ 8 -5 and
 426 ~ 3 Ma, based on increased black carbon accumulation (whose transportation to the sea
 427 would be favored by stronger winter winds) (Jia et al., 2003), planktic and benthic $\delta^{18}\text{O}$
 428 records and mixed layer temperatures (Holbourn et al., 2021, 2018), and clay mineralogy
 429 and grain size analysis (Wan et al., 2007). In the Sea of Japan, clay mineral assemblages
 430 and isotopic analyses of the silicate fraction covering the last 15 Ma suggest a step-wise
 431 drying of Central Asia at 12, 8 and 3.5 Ma (Shen et al., 2017). A sea surface temperature
 432 drop of ~ 8 °C recorded between 8-6.6 Ma (IODP Site U1425) is further interpreted as a
 433 sign for a marked EAWM intensification (Matsuzaki et al., 2020). At approximately the
 434 same time (8-4 Ma), an analysis of aluminosilicate end-member contributions shows a shift
 435 from a Taklimakan source to a Gobi-Chinese Loess Plateau source in dust exported to the
 436 Sea of Japan (Site U1430), suggesting changes in winter wind circulation patterns, or in the
 437 extent of these deserts (Anderson et al., 2020). Further east, in the North Pacific, a relative
 438 increase in Asian dust is also recorded at ~ 10 and ~ 4 Ma (D. Rea et al., 1985; Pettke et
 439 al., 2002).

440 2.3 Assessing monsoon forcing factors through climate models

441 The major drivers of long-term monsoonal variability are topography, land-sea distri-
 442 bution and global climate variations. Attributing a specific change to one or several of these
 443 factors is nevertheless complicated because they may co-vary, produce non-linear responses
 444 with potentially opposite effects, and are sometimes poorly dated or quantified. Modeling
 445 experiments, initially utilizing present-day geographies and boundary conditions, have en-
 446 abled a better understanding of the role played by these different forcing factors in shaping
 447 the regional monsoons.

448 2.3.1 Topographic influences on the South and East Asian monsoon with 449 present-day geography

450 The TP has long been considered instrumental in driving the Asian monsoonal cir-
 451 culation and precipitation, due to its elevated heated surface in summer acting as a "heat
 452 pump" driving the convergence of the surrounding air masses, in particular the ITCZ con-
 453 veying moisture from the Indian Ocean (Kutzbach et al., 1993; Zoura et al., 2019; Molnar
 454 et al., 1993; G. Wu et al., 2012). While the presence of the TP would indeed play a key role
 455 in the EASM establishment, with the heating of the Plateau inducing a cyclonic circulation
 456 anomaly promoting the advection of moist air masses towards Eastern Asia (Z. Zhang et
 457 al., 2007; H. Tang et al., 2013; R. Zhang, Jiang, & Zhang, 2017; G. Wu et al., 2012), several
 458 studies suggest that the orographic thermal forcing of the Plateau is not as important as
 459 previously thought for the SAM establishment. Instead, the physical barrier created by the
 460 topography would insulate the high moist static energy air (warm wet air masses) from the
 461 tropics from the low moist static energy air (cold dry air) brought by the westerlies and
 462 induces abundant orographic precipitations in northern India, through steering and forced
 463 ascension of moist air masses (Boos & Kuang, 2010, 2013; Acosta & Huber, 2020). Al-
 464 though the debate between a thermal versus mechanical role for TP on atmospheric flow

465 is still ongoing, a consensus emerged for the need of having an energetic-based framework
466 to study monsoon mechanisms at all timescales (Biasutti et al., 2018), and integrate them
467 within the planetary scale circulation dynamics (Rodwell & Hoskins, 1996). The condensa-
468 tion of moisture in these ascending air masses leads to important latent heat release in the
469 high troposphere over northern India, which acts as a positive feedback sustaining summer
470 convection and precipitation in South Asia (He, 2017). On the other hand, the large-scale
471 advection of the ITCZ over Asia in late spring and early summer is essentially driven by
472 surface ocean-continent temperature gradients and resulting pressure patterns (Acosta &
473 Huber, 2020; Merlis et al., 2013). Additionally, numerical simulations show that such en-
474 hanced migration of the summer ITCZ is scaled to Asian continental extent, and was likely
475 weaker in the past, when conditions of higher sea level and the presence of the Paratethys Sea
476 prevailed and translated into surface temperature gradients different from today (Z. Zhang
477 et al., 2007).

478 Smaller orographic features have also been shown to play a critical role in shaping and
479 strengthening both the SAM and EAM. The East African highlands, positioned in the path
480 of the ITCZ flow, impact both African and South Asian climate (Bannon, 1979; Rodwell
481 & Hoskins, 1995; Sepulchre et al., 2006). Recent studies have specifically demonstrated
482 that the East African highlands contribute greatly to the strengthening and concentration
483 of the Somali Jet, but that their absence would result, somewhat counter-intuitively, in
484 higher summer precipitation over the SAM region due to increased advection directly from
485 the ocean to the continent (H.-H. Wei & Bordoni, 2016; Chakraborty et al., 2009). This
486 last aspect hints that strong upwelling in the Arabian Sea, related to a strong Somali
487 Jet, may not always be positively correlated with heavy rainfall on land. The Anatolian-
488 Iranian landforms, by channeling the Somali Jet and insulating it from the subtropical
489 dry westerly flow, greatly contribute to enhancing moisture transport towards India and
490 eastern Asia (Acosta & Huber, 2020; He, 2017; H. Tang et al., 2013). Additionally, the
491 orographic precipitation observed over smaller Asian landforms, for example the Indian
492 Ghats or the Indo-Burman ranges, have been shown to promote moisture advection and
493 enhanced precipitation in adjacent regions located windward, especially over the Arabian
494 Sea, Bay of Bengal and South China Sea (S.-P. Xie et al., 2006).

495 The effect of the Pamir, Tian Shan and northeastern TP uplifts on the SAM remains
496 unclear. When these landforms are uplifted together with the Mongolian Plateau, a decrease
497 in precipitation is simulated over India (H. Tang et al., 2013). On the other hand, increased
498 rainfall is simulated in India while testing Pamir, Tian Shan and northeastern TP uplift but
499 maintaining a low Mongolian Plateau (R. Zhang, Jiang, & Zhang, 2017; R. Zhang, Jiang,
500 Zhang, Cheng, & Zhang, 2017), highlighting potential counter effects of those mountains
501 on the Asian climate. The Pamir, Tian Shan and Mongolian landforms are also perfectly
502 situated in the mid-latitudes to interfere both dynamically and thermally with planetary-
503 scale atmospheric circulation. Numerical simulations show that their uplift enhances aridity
504 in inland Asia (modern Gobi and Taklimakan deserts) and reinforces the EAWM circulation
505 patterns (X. Liu & Yin, 2002; X. Liu, Sun, et al., 2015; Baldwin & Vecchi, 2016; Sato, 2009;
506 R. Zhang, Jiang, Zhang, Cheng, & Zhang, 2017). By creating a cold pool and deviating
507 the westerlies towards northern Siberia, the Mongolian orography has been proposed to be
508 responsible for the northward migration of the subtropical Jet Stream in summer (White et
509 al., 2017; Shi et al., 2015; Sha et al., 2020), and for most of the Siberian High location and
510 intensity in winter (Sha et al., 2015; R. Zhang, Jiang, Zhang, Cheng, & Zhang, 2017).

511 *2.3.2 Modeling experiments using paleogeographic reconstructions*

512 Modeling studies based on present-day geographies are limited in the extent to which
513 they allow us to understand past changes, especially for periods when very different ge-
514 ographies prevailed. Paleoclimate simulations using presumed "realistic" paleo-boundary
515 conditions are however rarer, because they usually entail high computational cost and re-
516 quire important background knowledge of past boundary conditions (paleogeography, CO₂,

517 solar constant, ice-sheet volume). Such modeling experiments confirm that increasing Asian
518 continentality during the Oligocene and the Miocene, either due to the Paratethys Sea re-
519 treat and/or global eustatic sea level fall, is a key driver of increased moisture advection
520 towards Asia in summer (Fluteau et al., 1999; Ramstein et al., 1997; Z. Zhang et al., 2014;
521 Sarr et al., 2022). Under Oligocene conditions, the uplift of the peripheral Tian Shan, Pamir
522 and northeastern portions of the TP has also been shown to promote inland Asian aridity,
523 in contrast to the sole uplift of the central TP (R. Zhang, Jiang, & Zhang, 2017). Under late
524 Eocene conditions, the latitude at which the proto TP uplifts significantly and non-linearly
525 impacts East Asian climate; an uplift in the tropics ($\sim 11^\circ\text{N}$) induces precipitation on the TP
526 foothills but aridification in most of China, whereas an uplift at its modern latitude ($\sim 26^\circ\text{N}$)
527 increases moisture advection to eastern China and aridification in inland Asia (R. Zhang et
528 al., 2018).

529 Paleoclimate modeling offers a contrasted view regarding Eocene climate, with some
530 studies that simulate the presence of monsoonal climates at that time (Huber & Goldner,
531 2012; Licht et al., 2014; Z. Zhang et al., 2022), and others that do not (Tardif et al.,
532 2020; Z. Zhang et al., 2012; R. Zhang et al., 2018). Discrepancy among simulations can be
533 attributed to many differences in model structure, resolution, parametrization and setting
534 of boundary conditions, among which the choice of paleogeographic reconstruction is a key
535 aspect. Indeed, many features of Eocene paleogeography remain highly controversial, such as
536 the height of the incipient Tibetan topography, the shape of the Indo-Asian collision zone,
537 the height of peripheral landforms, or the land-sea distribution. Paleoclimate modeling
538 has nevertheless emphasized the importance of geography, rather than $p\text{CO}_2$ variations
539 or far-field ice-sheets effects, as the main driver of both SAM and EAM establishment and
540 intensification on long time scales (Farnsworth et al., 2019; Thomson et al., 2021). Accurate
541 paleogeographic constraints as well as a higher number of sensitivity experiments in a paleo
542 framework are therefore needed, and this is what we intend to contribute with this study.

543 **2.4 Summary: a diachronous establishment of SAM and EAM seasonal** 544 **features, the influences of geography and topography and remaining** 545 **uncertainties**

546 The past decade has welcomed a large amount of new data based on fossil material
547 and sedimentary archives, both onshore and offshore. The onset of the SAM, EASM and
548 EAWM and their evolution are now clearly shown to be diachronous and is tightly linked to
549 the evolution of land-sea distribution, topography and global climate over long time scales.
550 Some questions, nevertheless, remain to be answered. First, it is unclear whether Paleogene
551 records of highly seasonal precipitation at low latitudes ($< 20^\circ\text{N}$) describe a monsoon in
552 the strict sense, given that no wind proxies currently exist for this period and that Eocene
553 paleoclimate modeling studies propose widely diverging scenarios. CLAMP paleobotani-
554 cal analyses provide interesting insights on this point. They generally show that, although
555 strong precipitation seasonality is likely present since the Paleogene in broad parts of South-
556 eastern Asia (Bhatia, Khan, et al., 2021; Spicer et al., 2017; Herman et al., 2017), leaves did
557 not develop the morphological features characteristic of plants growing in present-day SAM
558 or EAM regions until the middle Miocene (~ 13 Ma) (Spicer et al., 2017; Bhatia, Srivas-
559 tava, et al., 2021). Instead, Paleogene Asian floras display a signature typical of the regions
560 dominated by the ITCZ seasonal migration, also referred to as the Indonesian-Australian
561 monsoon. The reorganization from an ITCZ-dominated to a monsoonal wind circulation
562 pattern may have initiated in the Oligocene (~ 28 - 23 Ma, Fig. 2), according to fossils from
563 northeastern India exhibiting a mixture of SAM and ITCZ leaf signatures (Spicer et al.,
564 2017). These typical monsoonal signatures in floras were likely driven by a combination
565 of both biotic and abiotic factors, although the CLAMP method is unable to disentangle
566 the forcing factors responsible for these specific signatures. They do, however, add to the
567 body of evidence suggesting that precipitation seasonality may have been established prior
568 to typical monsoonal wind circulation.

569 Second, although modeling studies have started to explore the effects of localized
 570 mountain ranges on Asian climate in a paleo context, there is plenty of room for improve-
 571 ment. For example, the climatic impact of these landforms have often been investigated
 572 simultaneously in a single climate simulation (e.g. the uplift of the Tian Shan and Mongo-
 573 lian orogens), though the main uplift episodes of these landforms may not be coeval (see
 574 section 3.1.2). Various hypotheses remain to be tested, such as identifying the drivers of the
 575 Siberian High development, and its impact on inland Asian aridity. Indeed, the presence of
 576 loess-like deposits since the Eocene in the Xining Basin were tentatively explained by the
 577 presence of an active Siberian High at that time (Meijer et al., 2021; Licht et al., 2014).
 578 This questions whether the Tian Shan and/or Mongolian orogens are as important for this
 579 wide anticyclone as predicted by modeling experiments, since these landforms are thought
 580 to have uplifted later, in the Oligocene to late Miocene (see section 3.1.2). In addition, the
 581 notion of elevation thresholds is a key point to tackle, as strongly non-linear responses of
 582 atmospheric circulation with respect to this parameter are expected. These considerations
 583 highlight the need for an accurate representation of Asian paleogeographic evolution in order
 584 to better understand monsoon history over long time-scales.

585 **3 Cenozoic paleogeographic evolution**

586 Although presenting a full literature review of paleogeographic evolution throughout
 587 the Cenozoic for the whole region of interest is outside the scope of this study, this section
 588 aims at providing an overview of the main features and persistent uncertainties regarding
 589 the paleogeographic evolution during the Cenozoic in our region of interest. This allows
 590 us to justify which sensitivity tests were performed on specific geographic units (Fig. 3),
 591 further introduced in section 4.2.1.

592 **3.1 Indo-Asia collision and Asian topography evolution**

593 The initiation of the Indo-Asian collision, marking the closure of the Neotethys Ocean,
 594 is dated around ~ 58 Ma based on sedimentological and paleomagnetic evidence (van Hins-
 595 bergen et al., 2019; Ingalls et al., 2016). The morphology of the *Greater India* portion (i.e.
 596 northern India) before the collision remains unclear. Depending on the collision scenario
 597 considered (see Kapp and DeCelles (2019); Poblete et al. (2021); J. R. Ali and Aitchison
 598 (2005) for reviews), it is represented as fully emerged (Ingalls et al., 2016; C. Wang et al.,
 599 2014), partially flooded or divided into two blocks separated by an oceanic basin until the
 600 middle-late Eocene (van Hinsbergen et al., 2012, 2019; W. Huang et al., 2015). From the
 601 Eocene onwards, the progressive migration of India to the north, its indentation into the
 602 Asian continent, and its counterclockwise rotation (Molnar et al., 2010) trigger widespread
 603 orographic changes in Asia, which are summarized below. The most important of these
 604 events is the TP and Himalayas uplift, and the growth of peripheral mountain ranges such
 605 as the Pamir, Tian Shan and the Mongolian Plateau.

606 **3.1.1 Himalayas-Tibetan Plateau uplift history**

607 The early uplift history of the Tibetan Plateau is highly debated. Oxygen isotope paleo-
 608 oaltimetry studies suggest that the Gangdese (Southern TP, Lhasa terrane) and Qiangtang
 609 mountains (Central TP) were already as high as ~ 4000 - 5000 m during the early Eocene
 610 (Rowley & Currie, 2006; L. Ding et al., 2014; Xu et al., 2013; Xiong et al., 2020; C. Wang et
 611 al., 2014). The robustness of this method is however questioned by isotope-enabled climate
 612 models, which demonstrate that factors other than elevation affect the $\delta^{18}\text{O}$ precipitation
 613 signature, including air mass provenance, climate, or water recycling (Poulsen & Jeffery,
 614 2011; Botsyun et al., 2016). A revised TP paleo-elevation of most likely lower than ~ 3000
 615 m in the Oligocene has therefore been suggested (Botsyun et al., 2019). The presence of
 616 frost-intolerant fossil flora assemblages in the central TP in the middle Eocene (T. Su et al.,
 617 2020) and late Oligocene (T. Su, Farnsworth, et al., 2019), as well as clumped isotope mea-

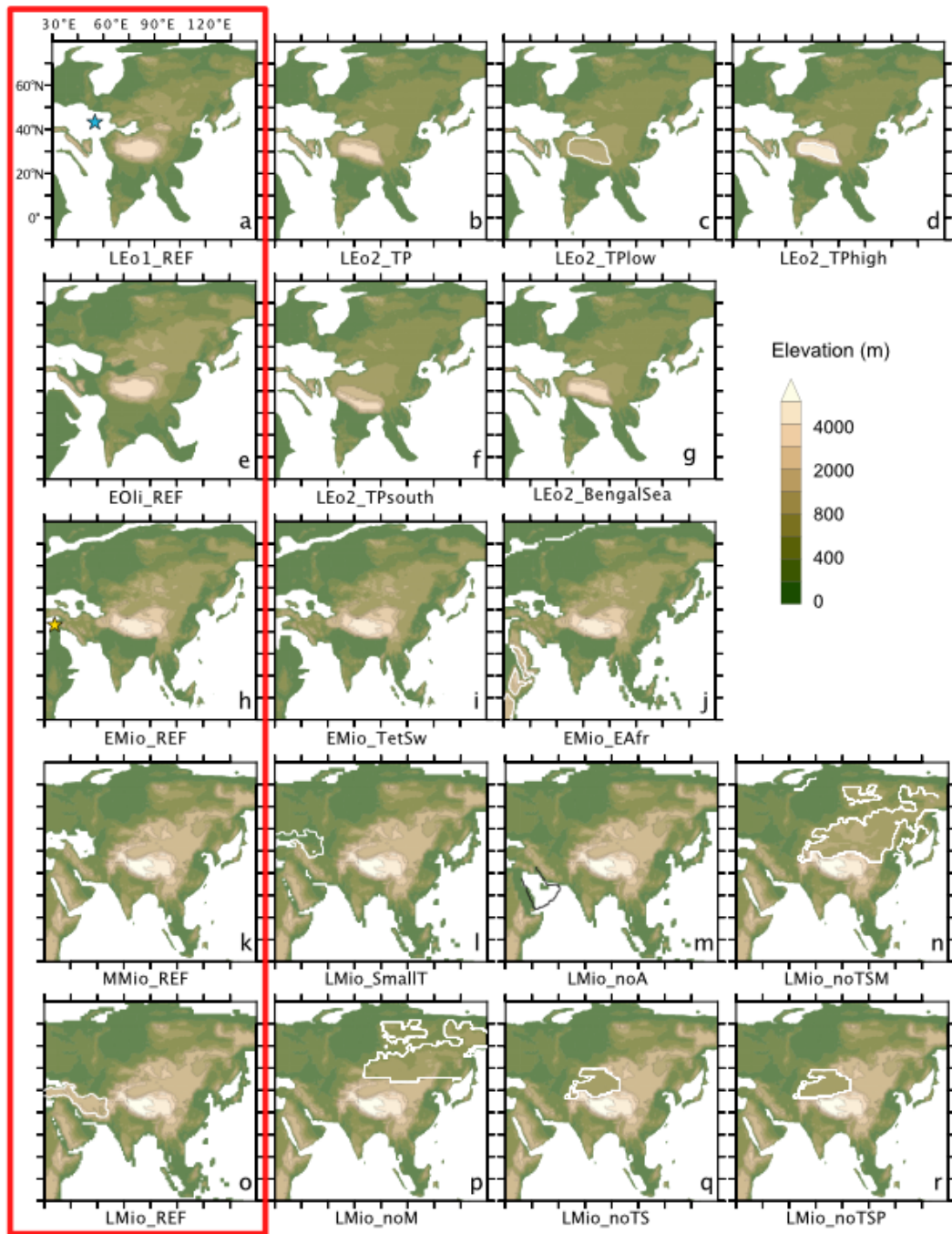


Figure 3. Paleogeographic configurations used in this study for the reference experiments (red box) and alternative configurations used for sensitivity tests. Blue star in (a) highlights the Paratethys Sea location, yellow star in (h) highlights the location of the Tethyan Seaway.

618 surements (Xiong et al., 2022) advocating for a subtropical climate in this region, also seem
 619 incompatible with a TP exceeding 4000 m. Therefore, a more complex morphology for this
 620 proto TP has been proposed, suggesting the presence of a low-elevation (~ 1500 m) valley
 621 dividing the almost fully uplifted (~ 3000 - 4000 m) Lhasa and Qiangtang terranes until the
 622 late Oligocene (T. Su, Farnsworth, et al., 2019; Xiong et al., 2020; Spicer et al., 2020; Xiong
 623 et al., 2022). In the periphery of the proto TP, data collected from different basins offer
 624 contradictory interpretations. In southeastern Tibet, at the extremity of the Qiangtang ter-
 625 rane, paleo-elevation calculated using the CLAMP method suggests that the Markam Basin
 626 was already high in the late Eocene (~ 3000 m) and could have reached its modern elevation
 627 (~ 3900 m) by the earliest Oligocene (T. Su, Spicer, et al., 2019). Alternately, paleo-elevation
 628 based on oxygen isotopic measurements of the nearby Jianchuan Basin (part of the same
 629 terrane) could not have exceeded ~ 1200 m in the late Eocene (Gourbet et al., 2017). The
 630 northern TP is proposed to have remained relatively low (< 2000 m) in the Eocene, before
 631 uplifting in the Oligocene or the early Miocene, based on isotopic paleoaltimetry from the
 632 Hoh Xil Basin (C. Wang et al., 2014). An early uplift of the northeastern TP has been
 633 proposed as early as the late Eocene by increased coniferous pollen proportion in the Xining
 634 Basin (Hoorn et al., 2012), although global cooling and the Paratethys Sea retreat may also
 635 have promoted coniferous trees expansion in this region, without uplift (Barbolini et al.,
 636 2020).

637 The history of the TP build-up from the Miocene onwards is less controversial. The
 638 Central TP is thought to have reached modern elevation by the early Miocene, while the
 639 northern TP uplift was completed by the late Miocene (C. Wang et al., 2014). Stratigraphy
 640 and detrital zircon analyses indicate that the western TP (Pamir-Karakoram-Hindu Kush)
 641 may have undergone several phases of uplift between the mid-Eocene to latest Miocene
 642 (Bershaw et al., 2012; L. Li et al., 2021; Blayney et al., 2016; Carrapa et al., 2015; D. Liu et
 643 al., 2017). Finally, the Himalayas start uplifting in the early Miocene (~ 20 Ma) and reach
 644 their modern elevation by the mid-late Miocene (~ 15 Ma) (Gébelin et al., 2013; L. Ding et
 645 al., 2017; Xu et al., 2018; C. Wang et al., 2014; Webb et al., 2017).

646 *3.1.2 Inland Asia and Myanmar mountain building*

647 Asian terranes located in the vicinity of the Himalaya-TP complex are impacted by the
 648 Indo-Asian collision. In Myanmar, the Indo-Burman ranges built up in a three-step process,
 649 in the late Eocene, at the Oligo-Miocene boundary and Pliocene respectively (Najman et
 650 al., 2020; Morley, 2018; Maurin & Rangin, 2009; Westerweel et al., 2020; Licht et al., 2018),
 651 as a result of the progressive indentation of the Burma block into the eastern Himalayan
 652 collision zone. The Sino-Burman Ranges, which inherit old landforms from previous colli-
 653 sions between Indochina and Sibumasu terranes (Metcalf, 2013), experienced major uplift
 654 in the mid-late Miocene (~ 13 - 9 Ma) (Cook & Royden, 2008; Clark et al., 2005), as a result
 655 of the eastward expansion of the Tibetan Plateau by crustal flow.

656 North of the TP, the collision propagates through reactivated pre-existing tectonic
 657 structures and triggers the uplift of the inland Asian orogenic belt: the Tian Shan and the
 658 Mongolian (Altai, Hangay and Sayan) landforms (Jolivet et al., 2010). The southwestern
 659 and Central Tian Shan uplift initiation is very loosely constrained, either in the middle to
 660 late Miocene (~ 15 Ma, (Käbner et al., 2016)) or during the late Oligocene to early Miocene
 661 (~ 25 Ma) (Bande et al., 2017; Macaulay et al., 2016, 2014; D. Liu et al., 2017). The
 662 northeastern Tian Shan, together with the Sayan and Altai regions (in western Mongolia),
 663 likely remained quiescent until the late Miocene (Bullen et al., 2003; Caves Rügenstein et al.,
 664 2017). After a pulse of uplift in the latest Miocene (~ 8 Ma) for the whole Tian Shan, Sayan
 665 and Altai mountains (W. Yuan et al., 2006; Caves Rügenstein et al., 2017; Charreau et al.,
 666 2005), regional landforms probably reached their modern elevation after a final Pliocene
 667 uplift phase (~ 5 - 3 Ma, (Bullen et al., 2003; Caves Rügenstein et al., 2014; De Grave et al.,
 668 2007, 2009; Jolivet et al., 2007; Vassallo et al., 2007)). The Hangay Dome uplift (in central
 669 Mongolia) began in the mid-Oligocene (Caves Rügenstein et al., 2014; Cunningham, 2001),

670 but the date of completion of the uplift remains debated, either in the mid-Miocene (~ 13
671 Ma, (Smith et al., 2016)), or the Pliocene (~ 3 Ma, (Yarmolyuk et al., 2008)).

672 **3.2 Evolution of the Peri-Tethys and Middle East region under the inter-** 673 **play of tectonics and global sea level changes**

674 Throughout the Cenozoic, sea level fluctuations in a context of tectonic convergence
675 have led to a profound paleogeographic reorganization and a general increase in continen-
676 tality of the regions encompassing the Mediterranean Sea, Middle East and inland Asia.

677 **3.2.1 Retreat of the Paratethys Sea and closure of the Neotethys Ocean**

678 During the Paleocene-Eocene, Eurasia was largely flooded by the proto Paratethys, an
679 epicontinental sea connected to the Arctic Ocean via the Turgai Strait and to the Neotethys
680 Ocean (Meulenkamp & Sissingh, 2003; Golonka, 2009). Paleoenvironmental analyses suggest
681 a maximal extension of the proto Paratethys Sea reaching the Tajik and Tarim Basins in
682 inland Asia at the Paleocene-Eocene transition (~ 56 Ma), followed by a long-term westward
683 retreat punctuated by marine transgressions of decreasing magnitude (Kaya et al., 2019;
684 Bosboom et al., 2014b, 2017; Carrapa et al., 2015). In Central Asia, the shrinkage of
685 the proto Paratethys Sea during the Eocene is attributed to far-field effects of the Indo-
686 Asia collision and to the early indentation of the Pamir and TP, leading to the progressive
687 infilling of the basins with sediments (Kaya et al., 2019; Carrapa et al., 2015; D. Liu et
688 al., 2017). The onset of the Antarctic ice-sheet at the EOT (~ 33.5 Ma) resulted in a
689 sea-level drop of ~ 70 m (Miller et al., 2020)), which drastically reduced the extent of the
690 Paratethys and favored its isolation (Kaya et al., 2019; Bosboom et al., 2014b, 2017). Water
691 exchanges between the Paratethys and neighboring oceans were progressively restricted, as
692 indicated by the increasing divergence between the $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ signatures of Paratethys
693 foraminifera shells and the other oceanic signals at that time (Ozsvárt et al., 2016). The
694 Tethyan Seaway, connecting the Indian and Atlantic Oceans, remained fully open before the
695 Miocene (Straume et al., 2020). ϵNd isotope data from foraminifers suggest that it became
696 strongly restricted around ~ 22 Ma (Bialik et al., 2019), with intermittent periods of closure,
697 as indicated by evidence of mammal exchanges between Eurasia and Africa across a land
698 bridge (Rögl, 1999; Harzhauser & Piller, 2007). The Tethyan Seaway permanently closed at
699 ~ 14 Ma, due to the combined effect of Arabian-Eurasian plate collision and glacio-eustatic
700 sea level fall associated with cooling following the Mid-Miocene Climatic Optimum (Bialik
701 et al., 2019). Until the late Miocene, the Paratethys Sea experienced phases of transgression,
702 extending as far as $\sim 60^\circ$ E, and regression that may have reduced its surface by up to 70%
703 (Palcu et al., 2021), thus highly modulating Eurasian continentality.

704 **3.2.2 Middle-East: Anatolian-Iranian Plateau, Zagros and Alborz Moun-** 705 **tains uplift**

706 During the Cenozoic, the Middle-East experienced multiple changes in land-sea distri-
707 bution and topography, forced by the collision of the Arabian plate with Eurasia that drove
708 the uplift of the Zagros Mountains and of the Iranian Plateau. The exact timing of the
709 Arabia-Eurasia collision in eastern Anatolia remains highly debated, with estimated dates
710 ranging from the Eocene-Oligocene (Karaođlan et al., 2016; Darin et al., 2018; McQuarrie
711 & Hinsbergen, 2013; Pirouz et al., 2017), the early Miocene (Okay et al., 2010; Gülyüz
712 et al., 2020), to the late Miocene (H. Su & Zhou, 2020; Z. Zhang et al., 2017). Most stud-
713 ies however propose an early Miocene age (~ 20 Ma) for the hard collision related to the
714 arrival of thick Arabian crust along the Bitilis suture zone (Okay et al., 2010; Cavazza
715 et al., 2018; Gülyüz et al., 2020) and a Miocene age for the uplift of most landforms (Zagros
716 and Alborz mountains, Anatolian-Iranian Plateau). After the collision, the Zagros oro-
717 gen builds up during three successive pulses occurring within the last ~ 20 Ma (Agard et
718 al., 2011; Mouthereau, 2011). Field geology, stable isotopes and thermochronology suggest

719 that the Alborz mountains uplift (north of the Zagros) occurred during the middle Miocene
 720 (~ 17.5 -13 Ma) (Ballato et al., 2010, 2015) and may have continued until the latest Miocene
 721 (Mouthereau et al., 2012). A Miocene age for the initiation of the eastern Anatolian-Iranian
 722 Plateau growth is also suggested, based on its youngest marine unit (~ 17 Ma) (Gülyüz
 723 et al., 2020) and apatite fission-track data (18-13 Ma) (Okay et al., 2010; Karaođlan et
 724 al., 2016). Stratigraphic evidence suggests an uplift phase of the eastern Anatolian-Iranian
 725 Plateau between 15 and 12 Ma due to crustal shortening and thickening (Mouthereau et al.,
 726 2012). In contrast, numerical simulations suggest that the final buildup of the Plateau is
 727 only reached during the late Miocene (~ 7 -10 Ma), likely sustained by slab break-off, mantle
 728 flow disruption and associated changes in dynamic topography, and followed by isostatic
 729 adjustment (François et al., 2014).

730 3.3 East Africa and Arabian Peninsula evolution

731 While much attention has been paid to the Eastern African topography evolution in the
 732 Neogene due to its implication for early hominins development, its history fits into a broader
 733 paleogeographic context rooted in the Paleogene (see Couvreur et al. (2021); Guillocheau et
 734 al. (2018) for a synthesis). Landforms in East Africa (Kenyan and Ethiopian plateaus) de-
 735 veloped during the late Cenozoic, triggered by a combination of tectonics, basaltic flooding,
 736 and large wave-length deformation related to the African superswell activity (Moucha &
 737 Forte, 2011; Faccenna et al., 2019; Roberts et al., 2012). Uplift in the Ethiopian region be-
 738 gan during the Eocene, with a doming event sustained by dynamic topography (Roberts et
 739 al., 2012; Faccenna et al., 2019) that then propagated to Kenya during the middle Miocene.
 740 In the early Oligocene, basaltic flooding in Ethiopia is thought to have contributed to the
 741 formation of these large-scale elevated features (Sembroni et al., 2016). Rifting initiated
 742 during the middle Miocene with the main uplift of rift shoulders (Wichura et al., 2015).
 743 Overall, past elevations during the Cenozoic are poorly constrained in East Africa, but
 744 modern elevations were most likely reached by the late Miocene to early Pliocene (Couvreur
 745 et al., 2021).

746 The Arabian Peninsula records successive marine and freshwater environments from
 747 the early Oligocene to the late Miocene, as indicated by paleo-ichthyofaunas (Otero & Gayet,
 748 2001). Before the Neogene, the Arabian plate is assumed to have been flat and at low
 749 elevation (Daradich et al., 2003) and thus particularly sensitive to sea-level changes. It was
 750 therefore largely submerged during the Eocene, and its southwestern part became partially
 751 emerged after the sea-level fall at the EOT (Barrier et al., 2018). During the Miocene, the
 752 Arabian plate was tilted when transiting over the East-Africa mantle plume (Vicente de
 753 Gouveia et al., 2018), leading to the uplift of the Red Sea margin, and the flooding of its
 754 northeastern corner, due to mantle convective drawdown (Daradich et al., 2003; Moucha
 755 & Forte, 2011). This mechanism promoted temporary transgression phases in the subdued
 756 northeastern region in early Miocene times (Barrier et al., 2018), until its full emergence
 757 in the late Miocene, possibly favored by global sea-level fall after the mid-Miocene Climate
 758 Transition (Golonka, 2009; Harzhauser & Piller, 2007; Miller et al., 2020).

759 4 Materials and Methods

760 4.1 IPSL-CM5A2 model

761 We perform simulations with the IPSL-CM5A2 Earth System Model (Sepulchre et
 762 al., 2020), an updated version of the IPSL-CM5A model (Dufresne et al., 2013) designed
 763 for deep-time paleoclimate simulations. It is composed of the atmospheric model LMDz5A
 764 (Hourdin et al., 2013), the land surface and vegetation model ORCHIDEE (Krinner et al.,
 765 2005) and the oceanic model NEMO (v3.6) (Madec, 2016) that also includes the LIM2 sea-
 766 ice model (Fichefet & Maqueda, 1997). The OASIS coupler ensures synchronization between
 767 the different model components (Valcke et al., 2006). The atmospheric model has a nominal
 768 horizontal resolution of 96x95 grid points (3.75° in longitude by 1.9° in latitude) with 39

769 irregularly distributed vertical levels from the surface to 40 km high. The ORCHIDEE model
 770 is coupled with LMDZ5a and redirects runoff water towards the ocean. The vegetation
 771 cover is represented through 11 Plant Functional Types, including one describing bare soil.
 772 NEMO solves ocean dynamics and thermodynamics equations on a tripolar grid, with two
 773 poles located below continental masses in the Northern Hemisphere, thereby avoiding North
 774 Pole singularity in the ocean (Madec & Imbard, 1996). It has a nominal resolution of 2° by
 775 2° refined up to 0.5° in the equatorial region, with 31 vertical levels, the thickness of which
 776 ranges from 10 meters near the surface to 500 meters in the deep ocean. Full description of
 777 IPSL-CM5A2 can be found in Sepulchre et al. (2020); Dufresne et al. (2013). IPSL-CM5A2
 778 has been previously used for paleoclimatic simulations of the Miocene (Burls et al., 2021;
 779 Sarr et al., 2022; Pillot et al., 2022) and the Eocene (Tardif et al., 2020, 2021; Toumoulin
 780 et al., 2022, 2020; Barbolini et al., 2020). The ability of the model to simulate the modern
 781 Asian climate has previously been described (Sepulchre et al., 2020; Tardif et al., 2020)
 782 and compared to observations (SI Fig. 14). While general atmospheric circulation patterns
 783 are well reproduced, IPSL-CM5A2 overall underestimates precipitation amounts and delays
 784 inland monsoonal precipitation onset by about a month, which are characteristics shared
 785 with the earlier IPSL-CM5A version of the model (Sepulchre et al., 2020).

786 4.2 Numerical simulations design

787 Simulations presented in this study are performed with paleogeographic configurations
 788 used in previous work, presented in Figure 3 and Table 1 and detailed below. The reference
 789 experiments have been run for 3000 years and the deep ocean has reached quasi-equilibrium
 790 after ~ 2000 years of integration. Sensitivity experiments are restarted from their reference
 791 experiment and run for 300 to 500 years until upper ocean equilibrium. Model outputs are
 792 averaged over the last 50 years of each simulation.

793 4.2.1 Paleogeographic reconstructions and sensitivity tests

794 Our late Eocene reference paleogeography (**LEo1_REF**, ~ 40 Ma, Fig. 3 a), previously
 795 introduced in Tardif et al. (2020), displays a fully emerged *Greater India* and a ~ 3000 m
 796 elevation TP. The proto Paratethys Sea fills the Tarim Basin, the Turgai Strait is closed and
 797 East African topography is low. Considering the ongoing debate surrounding the extension
 798 of India and TP morphology at that time, other configurations are tested: **LEo2_TP** (Fig. 3
 799 b) is an alternative late Eocene paleogeography, with a slightly different TP shape obtained
 800 with the "double collision model" (Poblete et al., 2021), from which sensitivity tests with
 801 a lower (~ 800 m, **LEo2_TP_{low}**, Fig. 3 c) and a higher TP (~ 4500 m, **LEo2_TP_{high}**,
 802 Fig. 3 d) are derived. Two additional tests are performed. The first exhibits a TP located
 803 more to the South (**LEo2_TP_{south}**, Fig. 3 f) and is obtained with the collision model
 804 from Jagoutz et al. (2015). The second features the presence of a remaining inlet between
 805 India and continental Asia (**LEo2_BengalSea**, Fig. 3 g), according to van Hinsbergen et al.
 806 (2012, 2019). The early Oligocene simulation (**EOli_REF**, ~ 33 Ma, previously introduced
 807 in Barbolini et al. (2020)) is designed to represent the Asian paleogeography right after the
 808 Eocene-Oligocene Transition (Fig. 3 e). The late Eocene paleogeography of LEo1_REF is
 809 used as a base, on which the sea level is eustatically lowered by 70 m (Miller et al., 2020),
 810 to account for a modern-sized ice sheet over Antarctica, as an end-member scenario. This
 811 leads to the emergence of continental portions in Arabia and North Africa and the drying
 812 of the Tarim Basin and lowlands north of the Turgai Strait, which all act to significantly
 813 increase Asian continentality. The former shallow connection between the Paratethys and
 814 the Tethys ceases in the Pamir-Zagros region at that time.

815 The early Miocene paleogeography (**EMio_REF**, ~ 20 Ma, Fig. 3 h), is from Poblete
 816 et al. (2021) and was previously introduced in Sarr et al. (2022) and Burls et al. (2021).
 817 It displays an Indian subcontinent translated 5° further north compared to the Eocene
 818 and an uplifted Central TP close to its modern elevation (~ 5000 m), whereas western and
 819 northeastern TP portions are lower (2000-3000 m). The Tethyan Seaway that previously

Table 1. List of reference simulations (bold) and sensitivity experiments used in this study. Abbreviations stand for East Antarctic Ice Sheet (EAIS), Antarctic Ice Sheet (AIS) and Greenland + Antarctic Ice Sheet (G+AIS). The last column details the paleogeographic configurations, highlighting the pairs of simulations that can be compared. For example, the effect of $p\text{CO}_2$ halving in the Eocene is obtained with the "LEo1_REF - LEo1_2X" anomaly.

Simulation	$p\text{CO}_2$	Ice sheets	Paleogeography
LEo1_REF	1120	-	from (Tardif et al., 2020)
LEo2_TP	1120	-	from (Poblete et al., 2021)
LEo2_TPlow	1120	-	from LEo2_TP, TP lowered to 800 m
LEo2_TPhigh	1120	-	from LEo2_TP, TP raised to 4500 m
LEo2_TPsouth	1120	-	from LEo2_TP, TP shifted to the South (Poblete et al., 2021)
LEo2_BengalSea	1120	-	from LEo2_TP, with a Bengal Sea (Poblete et al., 2021)
LEo1_2X	560	-	from LEo1_REF, $p\text{CO}_2$ halved to 560 ppm
EOli_REF	560	AIS	from LEo1_REF, 70 m sea level drop (Barbolini et al., 2020)
EMio_REF	560	EAIS	from (Poblete et al., 2021)
EMio_TetSw	560	EAIS	from EMio_REF, open Tethyan Seaway (120 m depth)
EMio_EAfr	560	EAIS	from EMio_REF, modern East African landforms
MMio_REF	560	AIS	from (Sarr et al., 2022)
LMio_smallT	560	AIS	from MMio_REF, reduced Paratethys
LMio_noAr	560	AIS	from LMio_smallT, immersed Arabia
LMio_noM	560	AIS	from LMio_smallT, Mongolia lowered to 800 m
LMio_noTS	560	AIS	from LMio_smallT, Tian Shan lowered to 800 m
LMio_noTSM	560	AIS	from LMio_smallT, Tian Shan+Mongolia lowered to 800 m
LMio_noTSP	560	AIS	from LMio_smallT, Tian Shan+Pamir lowered to 800 m
LMio_REF	560	G+AIS	from (Sarr et al., 2022)
LMio_1.5X	420	G+AIS	from LMio_REF, $p\text{CO}_2$ lowered to 420 ppm

820 connected the Mediterranean Sea and the Indian Ocean is closed, but a shallow sea persists
 821 at the location of the Arabian Peninsula. The alternative configuration **EMio_TetSw**, with
 822 a 120 m deep Tethyan seaway, is tested (Fig. 3 i). Considering the large uncertainties related
 823 to East African elevation, we also test the impact of a full uplift of this region as early as
 824 the early Miocene with the **EMio_EAfr** experiment (Fig. 3 j).

825 The mid and late Miocene paleogeographies are based on the PRISM4 reconstruction
 826 (Dowsett et al., 2016) used in PlioMIP2 (Haywood et al., 2020) with manual modifica-
 827 tions including the Australian continent shifted southward compared to present day and an
 828 emerged Sunda Shelf (Torsvik et al., 2008; Hall et al., 2012) (see Sarr et al. (2022) for details).
 829 In the middle (**MMio_REF**, ~12 Ma, Fig. 3 k) and late Miocene (**LMio_REF**, ~8 Ma, Fig.
 830 3 o) reference paleogeographic configurations, the Himalayas, Tibetan Plateau, Myanmar,
 831 inland Asia and East African landforms have reached their present-day configuration (Fig.
 832 3). In the mid-Miocene, the elevation of the Anatolian-Iranian orogen is half (~1000 m) that
 833 of present-day and the Paratethys Sea extends up to 60°E (**MMio_REF**), while by the late
 834 Miocene (**LMio_REF**), the Paratethys is strongly reduced and the Anatolian-Iranian land-
 835 forms have reached their modern elevation. An intermediate mid-late Miocene reconstruc-
 836 tion is tested, in which the impact of a Paratethys retreat is investigated (**LMio_smallT**,
 837 Fig. 3 l). Sensitivity tests are also performed on inland Asian landforms, by lowering to
 838 ~800 m the Mongolian Plateau (**LMio_noM**, Fig. 3 p), Tian Shan (**LMio_noTS**, Fig. 3
 839 q), Mongolia and Tian Shan (**LMio_noTSM**, Fig. 3 n), and finally, Tian Shan and Pamir
 840 (**LMio_noTSP**, Fig. 3 r). Additionally, **LMio_noAr** (Fig. 3 m), displaying an almost
 841 fully flooded Arabian Peninsula, is designed to account for the high sensitivity of this region
 842 to sea level fluctuations during that period. Note that **LMio_smallT** is the paleogeography
 843 to which all other late and middle Miocene simulations can be compared to.

844 4.2.2 pCO_2 , ice-sheets, solar constant and vegetation

845 CO_2 concentrations vary from 1120 ppm in the late Eocene to 560 ppm in the late
 846 Miocene (Table 1) (Foster et al., 2017; Rae et al., 2021) and other greenhouse gases are
 847 maintained at pre-industrial values. Acknowledging uncertainties regarding precise pCO_2
 848 concentrations during the Cenozoic, we perform two additional sensitivity tests on this
 849 parameter by prescribing 560 ppm in a late Eocene simulation (**LEo1.2X**) and 420 ppm
 850 in a late Miocene simulation (**LMio.1.5X**). Late Eocene experiments are considered ice-
 851 free, as pCO_2 is above the threshold for permanent Antarctic glaciation (Ladant et al.,
 852 2014; DeConto & Pollard, 2003). Unipolar glaciation over the Antarctic continent is used
 853 for the Oligocene, early and mid-Miocene, and both Greenland and Antarctic ice-sheets
 854 are present in late Miocene configurations (Bierman et al., 2016). The solar constant is
 855 adjusted (Gough, 1981) and set to 1360.19 W.m^{-2} for late Eocene and early Oligocene
 856 simulations, 1362.92 W.m^{-2} for early Miocene simulations and 1364.30 W.m^{-2} for late and
 857 middle Miocene simulations. Although orbital variations are known to impact monsoonal
 858 circulation (Zhisheng et al., 2015; P. Wang et al., 2005; Tardif et al., 2021), we prescribe
 859 a modern-like orbital configuration in all simulations because this study focuses on the
 860 impact of paleogeography. In the absence of congruent global vegetation reconstructions over
 861 the Cenozoic, all the experiments are forced with prescribed latitudinal bands of idealized
 862 vegetation cover based on modern distribution, as done in previous studies (Tardif et al.,
 863 2020; Laugié et al., 2020).

864 4.3 Monsoon metrics

The evolution of Asian monsoons over time is assessed through specific markers that
 are characteristic of the present-day SAM and EAM, such as precipitation amounts, season-
 ality, and wind direction and strength. The evolution of precipitation seasonality using the
 Monsoon Precipitation Index (MPI) (B. Wang & Ding, 2008) is expressed as follows:

$$\text{MPI} = \frac{\text{Seasonal range of precipitation}}{\text{Mean annual precipitation}}$$

865 where the seasonal range of precipitation is the difference between May to September minus
 866 November to March precipitations. Monsoon-like seasonality is considered to occur if the
 867 seasonal range is superior to 300 mm and more than half of the precipitation falls during the
 868 extended summer season ($\text{MPI} > 0.5$). This metric has the benefit of proposing a discrete
 869 metric for seasonality amplitude (from 0 to 1, instead of "with or without" monsoon). It also
 870 imposes that the wet season occur in summer, unlike other existing precipitation metrics
 871 (such as the ratio of the 3 wettest month over the 3 driest month precipitation, commonly
 872 used in paleobotany studies).

873 The wind patterns, described via their shape and strength, are the second fundamental
 874 characteristic that we focus on. First, we track the latitudinal migration of the Somali Jet
 875 over the Arabian Sea (Fluteau et al., 1999) as a proxy for SASM intensity. Second, we use the
 876 Webster-Yang Index (Webster et al., 1998), calculated as the difference between the zonal
 877 component of 850 and 200 hPa winds ($U_{850} - U_{200}$) in the region $0^\circ:20^\circ\text{N}$ to $40^\circ\text{E}:110^\circ\text{E}$, to
 878 evaluate the wind shear between the low and high troposphere over India, which describes the
 879 vigor of the SASM zonal circulation. Third, we use the positive thermal anomaly observed
 880 at 300 hPa over northern India in summer as a marker of the deep convection characterizing
 881 the modern SASM (Boos & Kuang, 2010; Acosta & Huber, 2020). This anomaly is due to
 882 the important latent heat release in the high troposphere provoked by the condensation of
 883 moisture in convecting air masses ascending the southern flank of the Himalayas (see Fig.
 884 14 d).

In addition to the precipitation seasonality, we explore the EASM evolution through
 the latitudinal displacement of the Jet Stream with respect to Tibetan and inland Asian
 landforms and with surface wind patterns. The EAWM is tracked principally with the
 evolution of temperature and pressure gradients, leading to the Siberian High formation
 in winter, and of its associated wind patterns. We also use the EAWM index (L. Wang
 & Chen, 2014), that accounts for both the North-South (Siberia-Maritime continent) and
 West-East (Siberia-North Pacific) pressure gradients and is expressed as follows:

$$I_{\text{EAWM}} = (2 \times \text{SLP}_{\text{S}} - \text{SLP}_{\text{NP}} - \text{SLP}_{\text{MC}}) / 2$$

885 where SLP_{S} , SLP_{NP} and SLP_{MC} is the area-averaged Sea Level Pressure (SLP) over Siberia
 886 ($40^\circ:60^\circ\text{N}$, $70^\circ:120^\circ\text{E}$), North Pacific ($30^\circ:50^\circ\text{N}$, $140^\circ\text{E}:170^\circ\text{W}$) and the Maritime continent
 887 ($20^\circ\text{S}:10^\circ\text{N}$, $110^\circ:160^\circ\text{E}$), respectively.

888 5 Results

889 For each regional monsoon, we first present the results obtained for the five reference
 890 numerical simulations (red box in Fig. 3). The underlying forcing factors are then explored,
 891 taking advantage of the sensitivity tests summarized in Table 1. The main outcomes in
 892 terms of seasonal winds and precipitations are qualitatively summed up in Table 2.

893 5.1 The South Asian monsoon, controlled by the orography surrounding 894 the Indian Ocean

895 5.1.1 Evolution of spatial and seasonal precipitation patterns

896 In the late Eocene, a wide arid region (< 500 mm/year) spreads from northern In-
 897 dia to north of proto Tibet (Fig. 4 a). The extent of this simulated arid region is most
 898 likely overestimated, which is further discussed in section 6.1. High mean annual precip-
 899 itation amounts (> 3000 mm/yr) are simulated over equatorial regions in southern India
 900 and southeastern Asia. There is no monsoon-like seasonality according to the MPI index in
 901 India, because the rainy season in this region occurs in winter (as previously described in
 902 Tardif et al. (2020)). The precipitation pattern evolves from the early Oligocene onwards,
 903 as the arid region progressively shifts to Arabia and northern Africa (Fig. 4 e, i, m, q).
 904 This allows the penetration of rainfall in summer into southeastern Asia and the onset of a

Table 2. Overview of each tested forcing factor on seasonal precipitation and wind circulation simulated in each region. Abbreviations stand for Transition Area (TA), cardinal directions (E, W, S, N), basins mentioned in Figure 1 (e.g. Tajik Basin, Tjk)

	SAM					EAM					Inland Asia		
	Precipitation			Winds		Precipitation			Winds		Precipitation		Precipitation
	Winter	Summer	Seasonality	Summer	Winter	Winter	Summer	Seasonality	Summer	Winter	Winter	Summer	Summer
No TP (LEo)	~	++				+	--	-	~	+	+(W)	~	~
High TP uplift (LEo)	~	+	~			++	+	-				-	~
TP South (LEo)	-	+				~	+	+				~	~
Bengal Sea (LEo)	+	++	-			+	+	~				+	+
CO2 decrease (EOT)	--	++	+(TA)	+		+	~	~	~	+	+	+	+
CO2 and sea level fall (EOT)	---	+++	+(TA, North. India)	++		--	++	+	+(E)	++(W/NW)	--	+	+
East Africa uplift (EMio)	+	++	-(India) + (TA, Himalaya)	+++ (Somali Jet)		+	+	-	+(SW)	+	+	+	-
Paratethys retreat (LMio)	~	++	+	++		+	+	~	+(SW)		--	--	~
Arabian peninsula emergence (LMio)	+	+	~	+(Somali Jet)		-	+	+	~	+(NW)	-	-	-
Iranian uplift (LMio)	--	++	+++	+++ (Somali Jet)		-	+	+			+	+	++
Mongolia uplift (LMio)	~	~	+			---	~	+++			+++ (NW)	+(orog.)	++(orog.) -- (Gobi, Takl., TS, Tjk)
TienShan uplift (LMio)	+	~	-			-	~	+	+(S)		+(orog.)	+(orog.)	+(orog.)
TienShan+Mongolia a uplift (LMio)	+	~	+			---	~	+++			+++ (NW)	+(orog.)	+(orog.) -(Takl. Tjk.)
TienShan+Pamir uplift (LMio)	+	+	--	+(Somali Jet)		~	~	~	+(SW)		+(orog.)	+(orog.)	++(orog.) -(Tjk)

monsoon-like seasonality in this region, as well as a reinforcement of monsoon-like seasonality in Eastern Asia. Over the course of the Miocene, as India drifts northwards, its southern tip receives less precipitation (~ 1500 mm/yr in the late Miocene, Fig. 4 q) and the intensity and extent of highly seasonal precipitation increases greatly over southern and eastern Asia (red dotted regions in Fig. 4 e, i, m, q). Orographic precipitation over the southern flank of the Himalayas initiates in the mid-Miocene simulation (~ 900 mm/yr, Fig. 4 m) and further increases in the late Miocene simulation (~ 1200 - 1500 mm/yr, Fig. 4 q). These changes in precipitation patterns in both space and time are triggered by a profound reorganization of surface pressure patterns and winds throughout the set of Cenozoic simulations.

5.1.2 Summer pressure and wind pattern reorganization

The summer (June to August, JJA) sea level pressure patterns in Eurasia in the late Eocene experiment (Fig. 4 b) contrast strongly with the modern ones (SI Fig. 14 f). A low-pressure center forms, driven by extreme summer temperature over southeastern Asia (up to 50 °C over *Greater India*, Fig. 4 d) whereas the Paratethys Sea and Neotethys Ocean are the locus of a wide high pressure cell (Fig. 4 b). This pressure pattern induces low tropospheric anticyclonic winds that counteract potential moisture advection from the Indian Ocean towards South Asian regions. From the early Oligocene onwards, the shrinkage of the Paratethys Sea and Neotethys Ocean increases Eurasian continentality and the surface of emerged land, such as Anatolia and Arabia. They become the locus of high summer temperatures (35 - 45 °C, Fig. 4 l, p, t), leading to the progressive widening of the South Asian low-pressure belt, while the Neotethyan anticyclone gradually regresses westward (Fig. 4 f, j, n, r). This low pressure cell reaches its modern-like extent in the mid-Miocene simulation (Fig. 4 n) but a modern-like magnitude (~ 14 hPa over Arabia and northern India) occurs only in the late Miocene simulation (Fig. 4 r, SI Fig. 14).

Mirroring this progressive strengthening and widening of the summer low-pressure belt over south Asia and Arabia, an increasing amount of moisture-loaded air masses are gradually advected from the Indian Ocean towards the continent over the Cenozoic. A weak cross equatorial flow (< 4 m/s, Fig. 4, against ~ 25 m/s today, SI Fig. 14), confined to low latitudes ($< 10^\circ$ N, Fig. 5 a, simulations LEo1_REF and EOli_REF), is simulated over the Arabian Sea in the late Eocene and early Oligocene experiments (Fig. 4 b, f). A mostly zonal proto-Somali Jet, also confined to low latitudes ($< 11^\circ$ N), is simulated in the early Miocene simulation (Fig. 5 b, simulation EMio_REF). From the mid-Miocene onwards, and following the reinforcement of the Asian low-pressure belt, the now fully-formed Somali Jet increases in strength and migrates further north over the Arabian Sea (~ 18 - 24° N) (Fig. 4 n, r, Fig. 5 b, simulations MMio_REF and LMio_REF). The simulated tropospheric circulation above southern Asia is drastically rearranged from the late Eocene to the Miocene simulation, as indicated by the Webster-Yang Index. This index doubles between the late Eocene and the early Oligocene (from ~ 5 to 11 m/s, Fig. 5 c, reference simulations LEo1_REF and EOli_REF) and keeps increasing in the Miocene (from ~ 16 to 21 m/s, Fig. 5 d, simulations EMio_REF, MMio_REF and LMio_REF). Finally, while a heating of the high troposphere (~ 300 hPa) due to deep convection is observed in most experiments (Fig. 4 d, h, l, p, t), this heat anomaly relocates over northern India only in the mid-Miocene experiment, following the increase in summer convective precipitation in this region driven by orographic uplift.

5.1.3 Limited effect of $p\text{CO}_2$ decrease and alternative Eocene Tibetan Plateau configurations on South Asian summer monsoon evolution

The non-linearity in the response of the Somali Jet through time calls for an individual analysis of the different forcing factors between each of these simulations. Our sensitivity experiments reveal the weak sensitivity of the SASM to $p\text{CO}_2$ changes in comparison to paleogeographic forcing. This is true for the late Eocene experiments, in which a change from 1120 ppm to 560 ppm (difference between experiments LEo1_2X and LEo1_REF) only induces minimal variations in the Webster-Yang Index, summer northward ITCZ migration

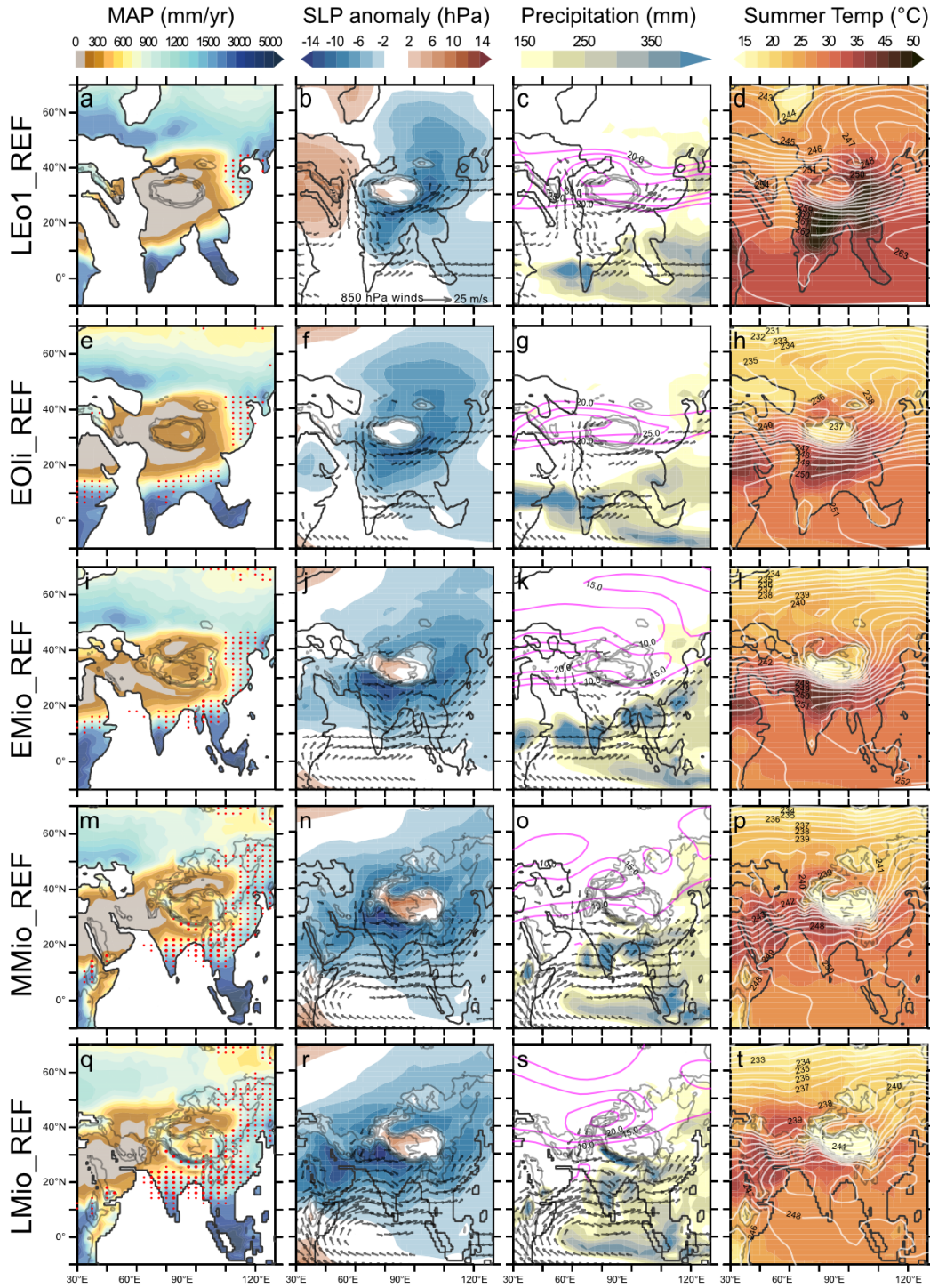


Figure 4. Evolution of summer monsoon diagnostics since the Eocene. **First column:** Mean Annual Precipitation (shading, mm/yr) overlain with regions where the Monsoon Precipitation Index is over 0.5 (thin red dots) and 0.75 (thick red dots); **Second column:** June-August (JJA) normalized Sea Level Pressure (SLP) anomaly defined as the difference between summer SLP and mean annual SLP (shading, hPa) and 850 hPa winds over 4 m/s (vectors); **Third column:** August Jet Stream speed (magenta contour, maximal zonal wind velocity in the mid to high troposphere, in m/s), August monthly precipitation amounts (shading, mm/month) and 850 hPa winds over 4 m/s (vectors) ; **Fourth column:** JJA mean temperature at 2 m (shading, Celsius) and 300 hPa atmospheric temperature (white contour, Kelvin). On all figure panels, topography is overlain as grey contours, at 1000 m intervals.

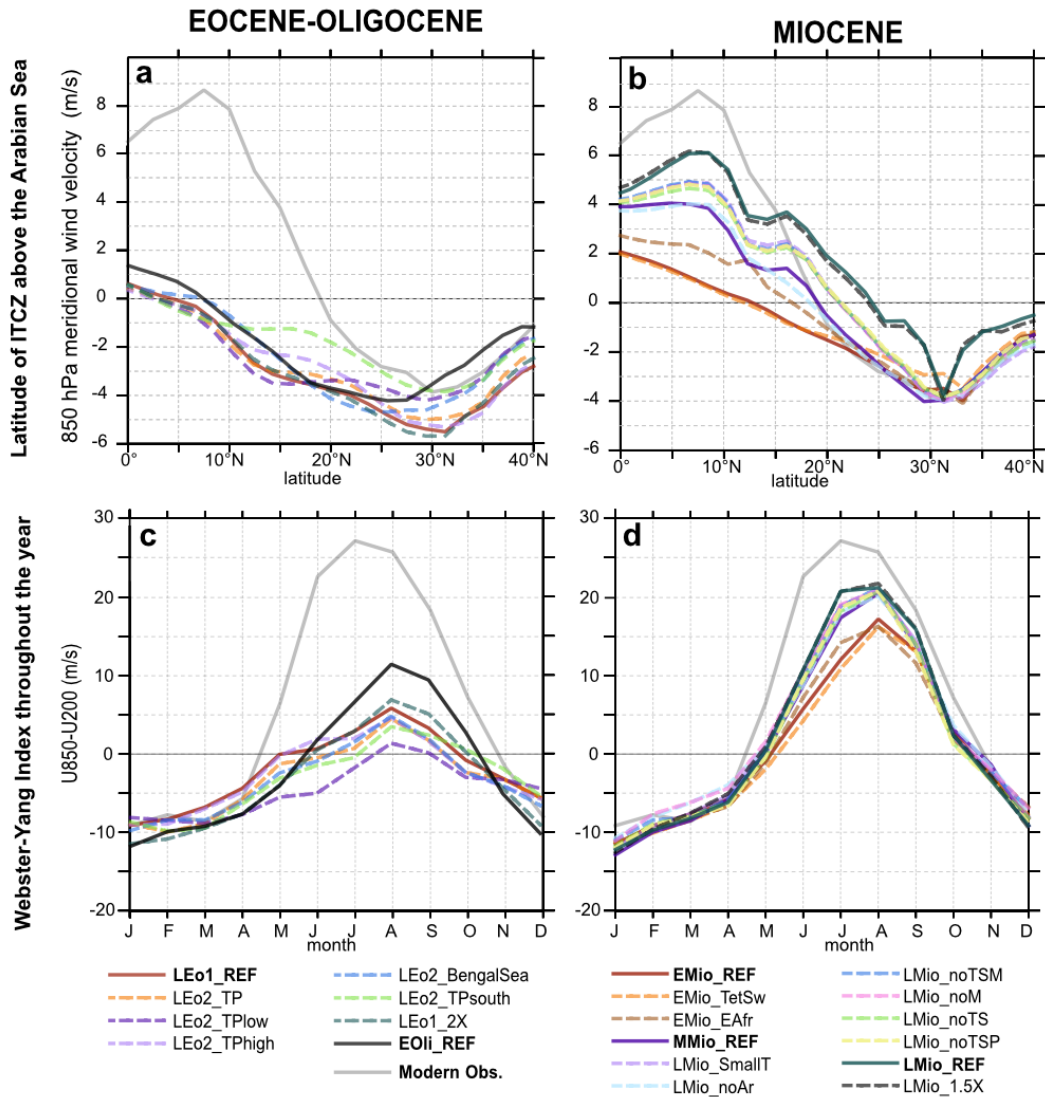


Figure 5. Estimation of the evolution of South Asian monsoon strength in all the simulations via: (a, b) the calculation of the summer (JJA) cross-equatorial flow latitude over the Arabian Sea, after Fluteau et al. (1999) and (c, d) the calculation of the Webster-Yang Index (WYI, 850 hPa minus 200 hPa zonal wind velocity, in the region 0°:20°N ; 40°:110°E, after Webster and Yang (1992)). For better readability, the experiments are split into Eocene-Oligocene (a, c) and Miocene (b, d) simulations. Reference simulations are highlighted in bold characters and continuous lines, while sensitivity tests are in dotted lines.

(Fig. 5 a, c), or monsoon-like seasonality extension (Fig. 7 f). This is also the case for the late Miocene experiments when the $p\text{CO}_2$ is changed from 560 ppm to 420 ppm (difference between experiments LMio_1.5X and LMio_REF in Fig. 5 b, d). The paleogeography of *Greater India* and Tibetan regions have been repeatedly proposed as a driver of monsoon onset or intensification (R. Zhang et al., 2012; Yu et al., 2018; Molnar et al., 1993). While alternative TP configurations lead to slightly different results, their impact on the simulated SASM large-scale wind circulation remains limited (Fig. 5 a, c, Fig. 6). In the late Eocene, increasing TP elevation strengthens the zonal wind shear over India (3000 m, Webster-Yang Index = ~ 6 m/s, for experiment LEO2_TP, Fig. 5 c) with respect to the Eocene experiment with low TP topography elevation (Webster-Yang Index = 1 m/s, LEO2_TPlow). Additionally, the magnitude of the Webster-Yang Index gradually increases with TP latitude, with values of 3 m/s for a southern edge at $\sim 20^\circ\text{N}$ (LEO2_TPsouth), 4 m/s at $\sim 25^\circ\text{N}$ (LEO2_TP), and 6 m/s at $\sim 30^\circ\text{N}$ (LEO1_REF). The presence of a sea inlet flooding *Greater India* in the late Eocene configuration limits the northward migration of the ITCZ in summer (from ~ 7 to 3°N between experiments LEO2_BengalSea and LEO2_TP, Fig. 5 c). It also produces a decrease in year-round precipitation, especially in south and inland Asia (Fig. 6 m, n, o).

5.1.4 *Continentality increase and East African and Iranian uplifts as major drivers of the South Asian monsoon*

The changes induced by the TP uplift or by its latitudinal location on South Asian climate are by far out-competed by the effects of land-sea distribution changes, or by the uplift of peripheral mountain ranges (Fig. 5). Indeed, the increase in continentality driven by the Eocene-Oligocene sea level fall induces colder and drier winters (Fig. 7 g), as well as warmer and wetter summers (Fig. 7 h). This results in a monsoon-like seasonality signal expanding over southeastern Asia (Fig. 7 i). It also translates into a doubling of the Webster-Yang Index (from 6 to 11 m/s between experiments EOLI_REF and LEO1_REF, Fig. 5 c) and an important reorganization of the summer Jet Stream behavior (see section 5.2.2). Likewise, increasing continentality during the Miocene appears to be key for inducing a more pronounced migration of the ITCZ over the Arabian Sea (from 18 to 21°N in the more continental case, Fig. 5 b), as well as a slight increase of the Webster-Yang Index (Fig. 5 d). This is achieved either with the emergence of the Arabian Platform (Fig. 8 h), or with the shrinkage of the Paratethys Sea (Fig. 8 e). In both cases, the amplification of the summer low pressure belt increases moisture advection from the Arabian Sea, leading to enhanced summer rainfall in the SAM domain. The increase in precipitation seasonality in the SAM domain is, however, more widespread with a Paratethys Sea retreat than with an emerged Arabian platform (Fig. 8 f, i), due to the contrasted effect of these paleogeographic changes on winter precipitation (Fig. 8 d, g). Both events are also shown to strongly impact central Asian aridity and the climate of the EAM domain.

Additionally, the buildup of both East African (difference between experiments EMio_EAfr-EMio_REF) and Anatolian-Iranian landforms (difference between experiments LMio_REF-LMio_smallT) is critical for the amplification of the ITCZ migration and strengthening of the SAM. Each uplift triggers a $\sim 4\text{-}5^\circ$ northward migration of the ITCZ over the Arabian Sea in summer, as well as a slight increase in the Webster-Yang Index in July (Fig. 5 b, d and SI Fig. 16). The uplift of East African landforms under Early Miocene conditions reinforces the Somali Jet and brings moisture to Iran and the SAM domain in summer (Fig. 9 e), which results in an intensification of monsoon-like seasonality in the Himalayan foothills and in southeastern Asia (mainly Myanmar, Fig. 9 f). Anatolian-Iranian landforms' final uplift under late Miocene conditions promotes winter aridity and summer precipitation in the SAM domain through a large strengthening of the Somali Jet, resulting in reinforced monsoon-like seasonality (Fig. 9 j, k, l). By preventing the advection of drier westerly mid-latitudes winds to lower latitudes, the Anatolian-Iranian Plateau channels humidity towards inland Asia in summer, reducing the aridity of the region. On the other hand, the sole effect of the Tethyan Seaway closure, without strong modification of land extension in the Arabian Peninsula region, remains limited in our experiments (+1 m/s for the Webster-Yang Index

1009 between experiments EMio_REF and EMio_TetSw in Fig. 5 d), in contrast to the hypoth-
 1010 esis that the closure of the Tethyan Seaway may contribute to altering the intensity of the
 1011 monsoon during the Miocene (Bialik et al., 2019, 2020; Rögl, 1997; J. Sun et al., 2021),

1012 **5.2 The East Asian monsoon under the combined influence of Asian orog-** 1013 **raphy, continentality and $p\text{CO}_2$**

1014 ***5.2.1 East Asian summer monsoon: precipitation patterns and summer*** 1015 ***Jet Stream position evolution***

1016 As opposed to the South Asian region, monsoon-like precipitation seasonality is sim-
 1017 ulated in eastern Asia in all reference experiments, although it remains initially confined
 1018 to northeastern China in the Eocene (Fig. 4 a). As this region of monsoon-like precipi-
 1019 tation expands to southeastern and inland China throughout the Cenozoic and reaches a
 1020 modern-like extent in the mid-Miocene experiment, the Paleogene broad zonal arid band
 1021 retreats from China (Fig. 4 m). Today, precipitation in eastern China is triggered by the
 1022 penetration of the Meiyu-Bayu front as it follows the migration of the Jet Stream north of
 1023 the TP during summer (Kong et al., 2017). In our simulations, the summer Jet Stream (ap-
 1024 proximated by the location of maximum zonal wind velocity in the troposphere in August)
 1025 circulates above the proto TP and eastern China in a strong zonal flux in the Eocene and
 1026 Oligocene experiments (up to 30-35 m/s centered at $\sim 30^\circ\text{N}$ Fig. 4 c, g). It then weakens
 1027 (10-20 m/s) and progressively migrates north of the TP in the Miocene experiments (Fig. 4
 1028 k, o, s). The Jet configuration evolves from purely zonal (east-west) to more SW-NE. This
 1029 relocation of the Jet Stream is accompanied by increased precipitation in eastern China,
 1030 southeastern Siberia and Japan, reaching 200-300 mm in August (Fig. 4 k, o, s), which
 1031 locally accounts for up to 25% of the annual precipitation. These results suggest that the
 1032 evolution of the Jet Stream seasonal displacement indeed amplifies the EASM rainy season,
 1033 although seasonal precipitation is simulated prior to the Jet acquiring its modern position
 1034 in summer.

1035 ***5.2.2 Summer Jet Stream migration driven by $p\text{CO}_2$ decrease at the EOT*** 1036 ***and Tian Shan uplift in the Miocene***

1037 Our sensitivity experiments allow us to identify the drivers responsible for the estab-
 1038 lishment of modern-like Jet Stream seasonal migrations over Asia. All 1120 ppm Eocene
 1039 simulations display a behavior of the Jet opposite to its modern counterpart, with a north-
 1040 ward migration in winter and a southward migration in summer independent of the TP
 1041 configuration (Fig. 10 a, Fig. 6 e, h, k, n). Halving $p\text{CO}_2$ at the EOT reverses this ten-
 1042 dency (LEo1.2X and Eoli_REF, compared to LEo1_REF), inducing a migration north of
 1043 the TP in late summer to early fall (Fig. 10 a, Fig. 7 e, h) and an important decrease in
 1044 winter zonal velocity (~ 50 m/s against ~ 70 m/s in other Eocene simulations, Fig. 10 a).
 1045 This northward migration is however restricted to the TP region and the Jet Stream remains
 1046 at low latitude over eastern Asia ($\sim 35^\circ\text{N}$, Fig. 10 b, SI Fig. 18), with low summer rainfall
 1047 amounts (100-150 mm/month). All subsequent Miocene simulations display an important
 1048 latitudinal migration of the Jet over the year, both in the TP region (Fig. 10 c, e) and over
 1049 eastern Asia (Fig. 10 d, f). The northward migration of the Jet is comparable between
 1050 early Oligocene and early Miocene experiments (up to $\sim 38-40^\circ\text{N}$ in September), despite a
 1051 higher (from 3000 to 4000 m) and wider TP, due to the incipient uplift of northeastern
 1052 and northwestern TP portions in the early Miocene simulation (the Tian Shan is ~ 1000 m
 1053 and the NE Tibet ~ 2000 m high). We therefore suggest that the Jet Stream latitudinal
 1054 migration is primarily driven by $p\text{CO}_2$ decrease and resulting changes in tropospheric tem-
 1055 perature gradients, and that a significant threshold in the behavior of the Jet Stream may
 1056 be crossed at the greenhouse-icehouse transition (EOT, ~ 34 Ma). In contrast, variations in
 1057 the Jet Stream latitudinal migration during the Miocene appear to be driven by topographic
 1058 changes. Indeed, sensitivity experiments show that the Tian Shan, and to a lesser extent
 1059 Mongolian and Pamir topographies, are critical to deviate the Jet northward in summer,

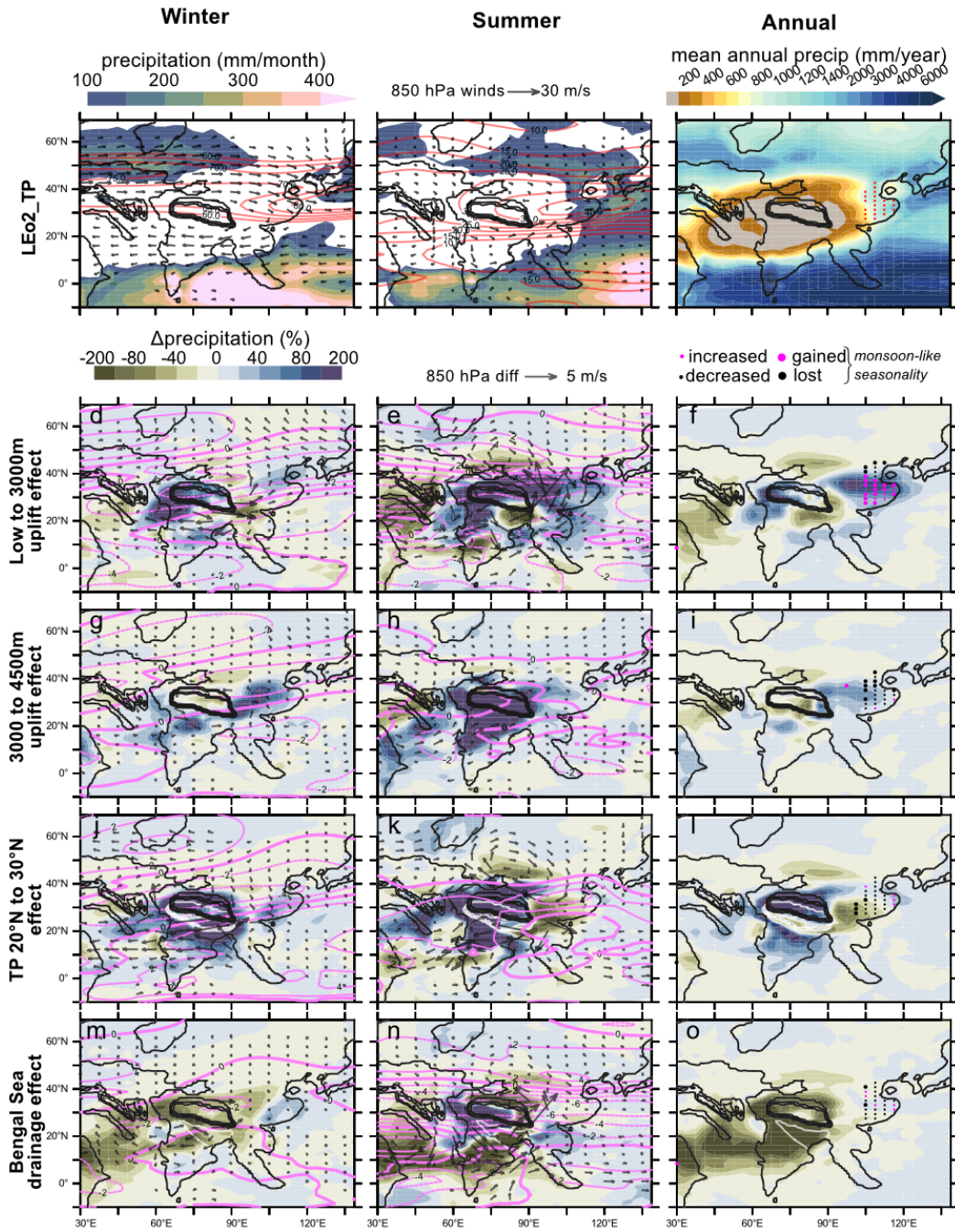


Figure 6. Sensitivity to the Tibetan configurations on Eocene climate: (a, d, g, j, m) winter precipitation (shading), 850 hPa winds (vectors), jet stream speed (contour) ; (b, e, h, k, n) on summer precipitation (shading), 850 hPa winds (vectors), jet stream speed (contour); (c, f, i, l, o) mean annual precipitation (shading) and monsoon-like seasonality according to the MPI index (red dots). Sensitivity tests are compared to the LEO2_TP simulation (a, b, c) and anomalies (d to o) are expressed as "test - LEO2_TP", except for the "lowered TP effect" which is "LEO2_TP - LEO2_TPflat"). Precipitation anomalies (shading) are expressed in % and normalized by the averaged precipitation of both simulations (therefore, a change from or to zero mm/year accounts for +/- 200%). In (f, i, l, o): four types of seasonality change induced by uplift are displayed (e.g., regions showing a decrease of monsoon-like seasonality after lowering the TP are in small black dots) ; white and black topographic contours indicate the LEO2_TP and tested topography.

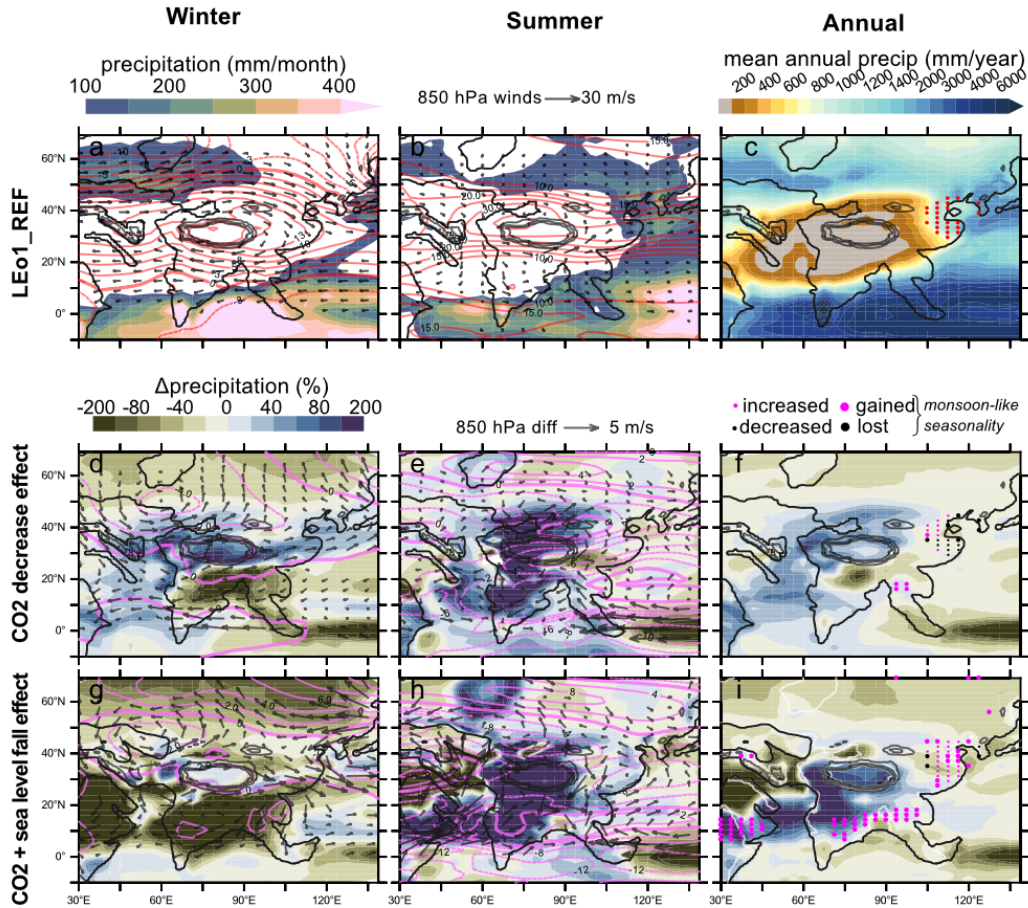


Figure 7. Influence of $p\text{CO}_2$ decrease and sea level fall at the EOT on: (a, d, g) winter precipitation (shading), 850 hPa winds (vectors), Sea Level Pressure (contour) ; (b, e, h) on summer precipitation (shading), 850 hPa wind (vectors), jet stream speed (contour); (c, f, i) mean annual precipitation (shading) and monsoon-like seasonality according to the MPI index (red dots). Sensitivity tests are compared to the LEO1_REF simulation (a, b, c) and anomalies (d to i) are expressed in order to highlight the impact of the test (i.e. test-LEO1_REF). We refer the reader to Figure 6 for an extended description of the panels.

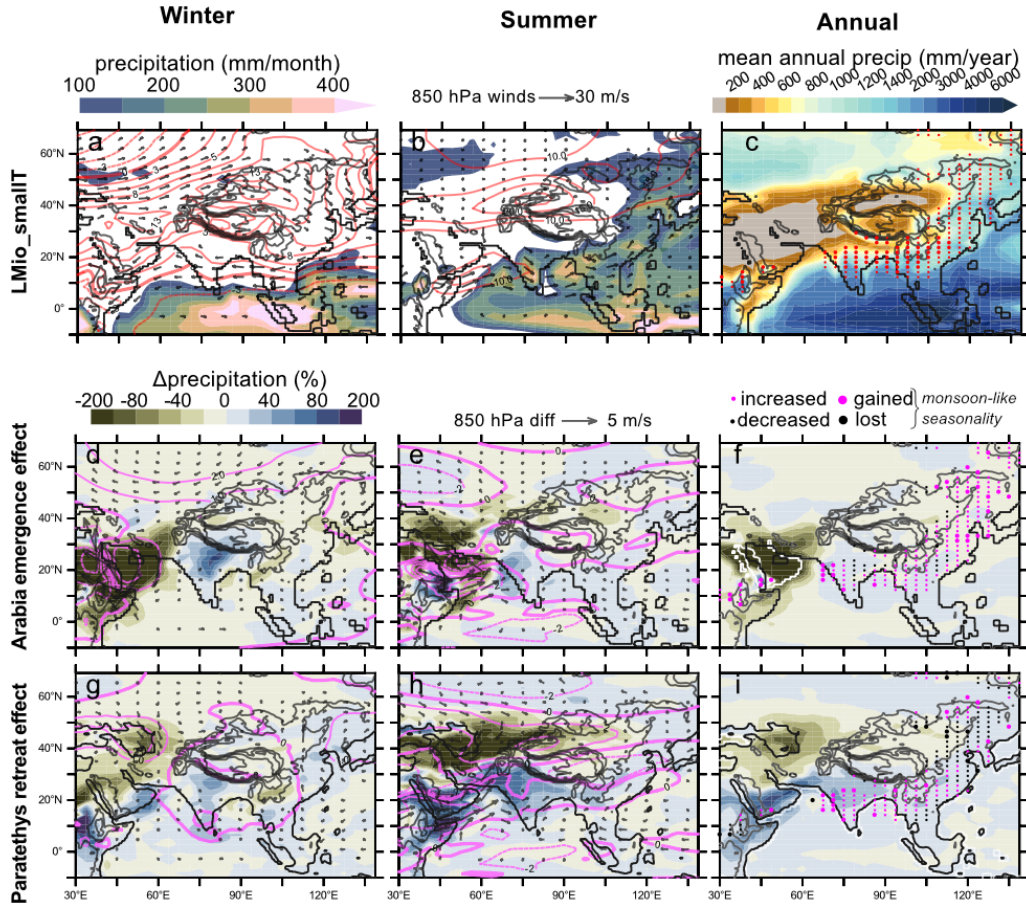


Figure 8. Influence of continentality increase in a mid-late Miocene context on: (a, d, g) winter precipitation (shading), 850 hPa winds (vectors), Sea Level Pressure (contour) ; (b, e, h) on summer precipitation (shading), 850 hPa winds (vectors), jet stream speed (contour); (c, f, i) and on mean annual precipitation (shading) and monsoon-like seasonality according to the MPI index (red dots). Sensitivity tests are compared to the LMio_smallT simulation (a, b, c) and anomalies (d to i) are expressed in order to highlight the impact of the continentality increase (i.e. LMio_smallT-test). We refer the reader to Fig. 6 for an extended description of the panels.

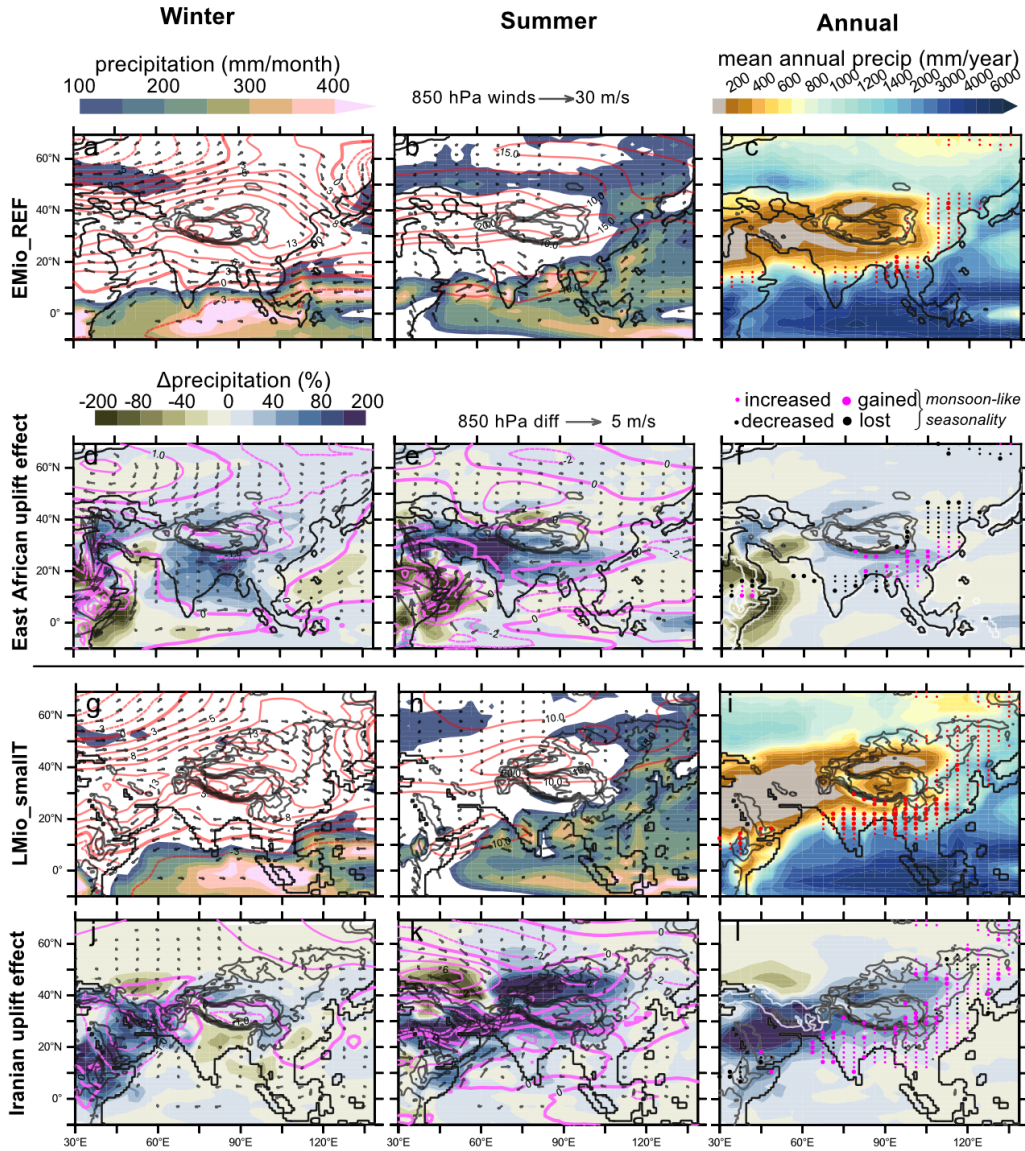


Figure 9. Influence of East African uplift in the early Miocene (a to f) and of Anatolian-Iranian uplift in the late Miocene (g to l) on: (a, d, g, j) winter precipitation (shading), 850 hPa winds (vectors), Sea Level Pressure (contour) ; (b, e, h, k) on summer precipitation (shading), 850 hPa winds (vectors), jet stream speed (contour); (c, f, i, l) and mean annual precipitation (shading) and monsoon-like seasonality according to the MPI index (red dots). East African (Anatolian-Iranian) uplift is compared to the EMio_REF simulation (LMio.SmallT) (a, b, c and g, h, i) and anomalies (d, e, f and j, k, l) are expressed in order to highlight the impact of uplift (i.e. EMio_EAfr-EMio_REF and LMio_REF-LMio_smallT respectively). We refer the reader to Figure 6 for an extended description of the panels.

1060 both in the TP region and in eastern Asia (from 39-41°N in LMio_noTS, LMio_noTSP,
 1061 LMio_noTSM and LMio_noM to 43°N in MMio_REF and LMio_REF, Fig. 10 e, Fig. 11 h,
 1062 k, n). Northward migration of the Jet Stream in summer induces orographic precipitation
 1063 on the uplifted landforms, aridification in the windward regions (-20 to -60% of summer pre-
 1064 cipitation in eastern Kazakhstan and Uzbekistan) and increased summer rainfall in eastern
 1065 Asia (Fig. 11 e, h, k, n). Monsoon-like seasonality in eastern Asia (magenta dots in Fig. 11
 1066 f, i, l, o) also greatly increases, due to the important effect of uplift of these mountains on
 1067 wintertime climate as well (see section 5.2.4). On the other hand, changes in the configura-
 1068 tion of the Tethyan Seaway in the early Miocene (comparing EMio_REF and EMio_TetSw),
 1069 the Anatolian-Iranian Plateau uplift (comparing LMio_SmallT and LMio_REF), or a $p\text{CO}_2$
 1070 decrease (comparing LMio_REF and LMio_1.5X) in the late Miocene, do not yield important
 1071 variations in the Jet Stream migration (Fig. 10 c to f).

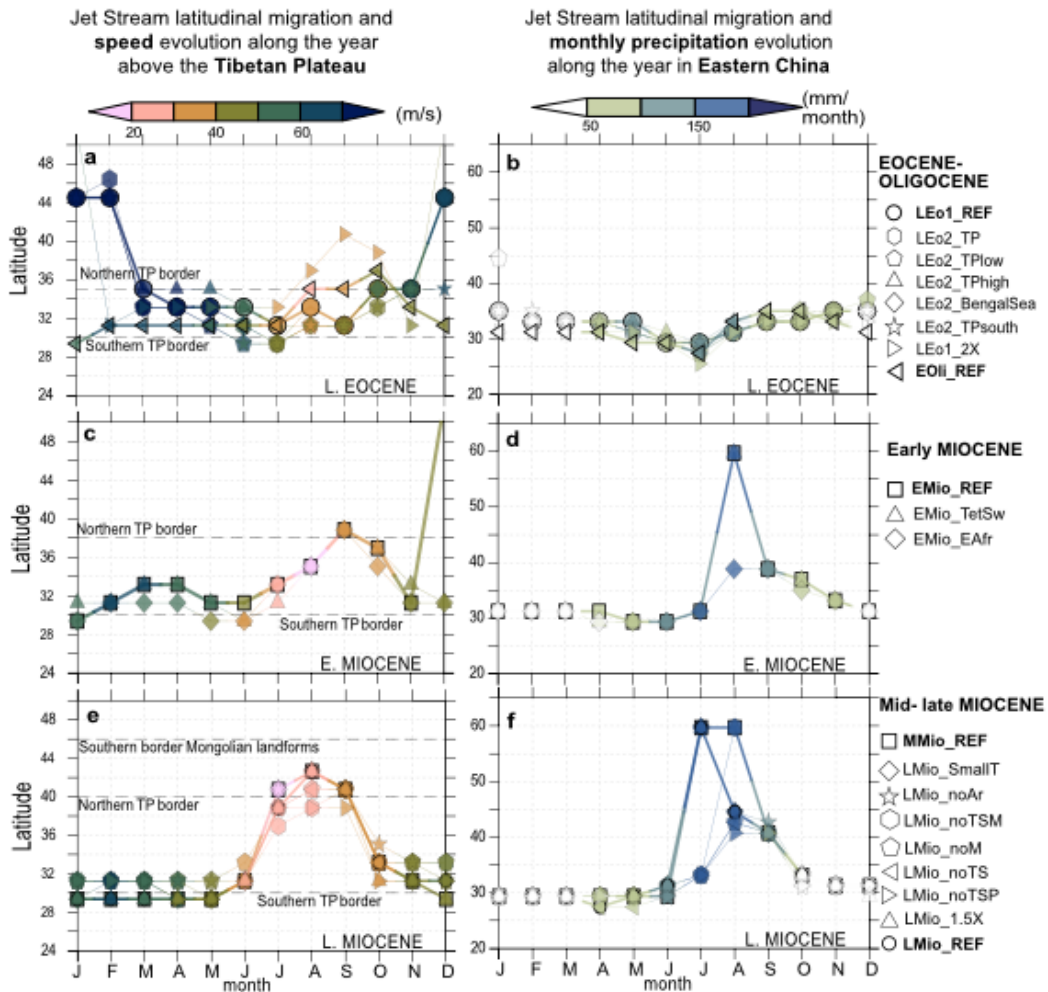


Figure 10. Jet Stream behavior throughout the year over the Tibetan Plateau and eastern China for all sensitivity experiments, divided into (**up**) late Eocene, (**mid**) early Miocene and (**bottom**) mid to late Miocene. (**left**) Jet Stream latitudinal migration and speed (shading) evolution along the year in the Tibetan region (averaged over 80:90°E). Mean latitudes of the northern and southern bounds of the Himalaya-Tibet and Mongolian landforms are indicated. (**right**) Jet Stream latitudinal position and precipitation amounts (shading) throughout the year in east Asia (averaged over 100:120°E).

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5.2.3 *East Asian winter monsoon: the Siberian High genesis*

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Our results suggest that pressure and wind patterns characterizing the modern EAWM established progressively, following an important reorganization of surface temperature gradients (Fig. 12, Fig. 13). In our Eocene experiment, a high-pressure band centered around 30°N is simulated but remains mainly zonal, following the large-scale winter descending branch of the Hadley cell driven by latitudinal surface temperature gradients (Fig. 12 b). The temperature difference between the Tibetan region (~ 10 °C, located below the high pressure cell) and neighboring North Pacific Ocean (~ 30 °C) remains relatively low (~ 20 °C, Fig. 12 a) compared to modern winter temperature difference (~ 40 °C, Fig. 14 h). As a result, no significant pressure difference is simulated between the Asian continent and the North Pacific in the Eocene and the calculated EAWM index is low (< 5 , Fig. 13), reflecting only the North-South pressure gradient between Asia and the Maritime Continent. The extent of the Asian high pressure zone increases gradually in the early Oligocene and early Miocene experiments (Fig. 12 d, f), but a modern-like Siberian High is only simulated in the mid and late Miocene experiments, when the anticyclone relocates over Siberia (~ 40 - 60 °N, Fig. 12 h, j). Additionally, the (North Pacific) Aleutian low relocates 10° to the south from the early Oligocene onwards and steadily weakens until the late Miocene (from over ~ 14 to 10 hPa anomaly). These changes concur to progressively increase both North-South and East-West pressure gradients, and therefore the EAWM index, which reaches maximal values in the mid and late Miocene experiments (~ 15 , Fig. 13). Following this simulated pressure pattern reorganization, low tropospheric winds circulating over China display a progressive change in orientation: while remaining primarily zonal in the Paleogene and early Miocene experiments (Fig. 12 b, d), they increasingly tilt in the meridional direction under the development of the Siberian High in the mid and late Miocene experiments (Fig. 12 h, j). This profound reorganization of atmospheric circulation is driven by temperature gradient changes and mainly due to the progressive cooling of the Tibetan, Mongolian and Siberian regions, as illustrated by the ~ 22 ° southward shift of the 0 °C isotherm between the late Eocene and the late Miocene simulations (Fig. 12 a, i).

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5.2.4 *Siberian High inception driven mainly by Mongolian uplift in the Miocene and its impact on precipitation seasonality*

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Our sensitivity experiments reveal the very strong impact of the Mongolian landform uplift (alone or together with the Tian Shan) in the mid-late Miocene on Asian winter pressure patterns. Indeed, their growth induces a strengthening of both the Siberian High (up to ~ 6 hPa, Fig. 11 d, g) and the Aleutian Low, resulting in a reinforcement of the East-West and North-South pressure gradients and ultimately of the EAWM index (from ~ 8 to ~ 15 , Fig. 13). Mongolian (and/or Tian Shan and/or Pamir) uplift also results in localized precipitation intercepted by the landforms (Fig. 11 d to o), despite the overall dryness of the inland Asian region (SI Fig. 15). As Mongolian uplift deviates the winter westerlies flux northward, Eastern China is swept by colder and drier winds coming from Siberia, which translates into a marked decrease in winter precipitation (up to 80% decrease, Fig. 11 d). These changes in the wintertime climate lead to an important expansion of regions displaying monsoon-like precipitation seasonality in eastern Asia (Fig. 11 f, i). The uplift of Tian Shan landforms, alone or together with the Pamir, has a lower impact on the winter pressure patterns compared to Mongolia uplift (Fig. 13) but nonetheless generates an almost complete drying on their eastern flanks (modern Gobi and Taklimakan deserts) because they act as an orographic barrier to westerly moisture flux (Fig. 11 g, j, m). Considering other sensitivity experiments, Asian winter climate exhibits only a weak response to changing Tibetan orography or $p\text{CO}_2$ values for Eocene conditions (SI Fig. 17 a to g, Fig. 13). Likewise, the Miocene winter pressure patterns in Asia are only mildly impacted by the Anatolian-Iranian landform uplift, the closure of the Tethyan Seaway, the Paratethys extent, or a $p\text{CO}_2$ drop from 560 to 420 ppm, when compared to the effect triggered by the Mongolian Plateau uplift (Fig. 13, SI Fig. 17 i to n and s, t).

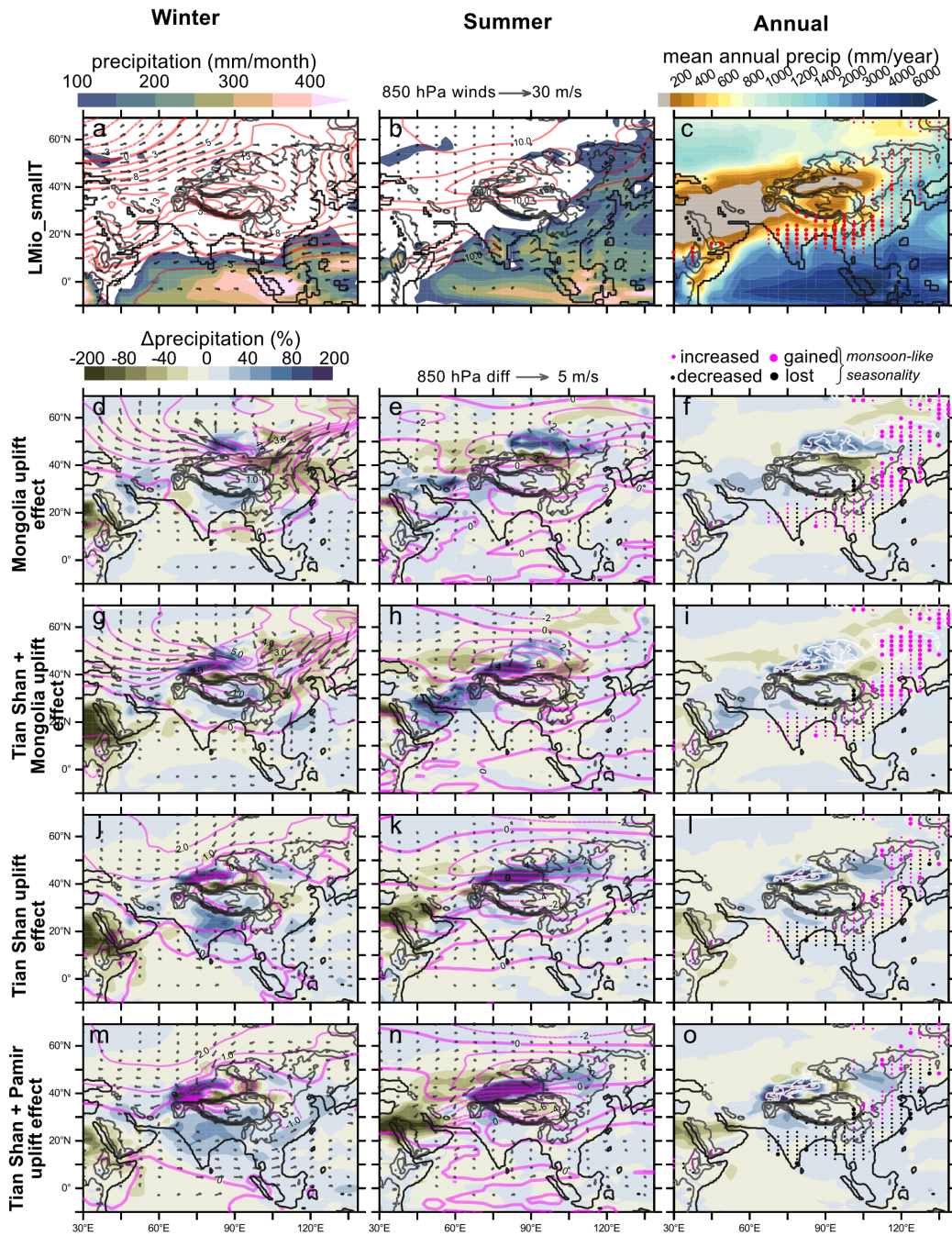


Figure 11. Influence of Mongolia, Tian Shan and Pamir regional uplifts on (left column) winter precipitation (shading), 850 hPa winds (vectors), Sea Level Pressure (contour) ; (middle column) on summer precipitation (shading), 850 hPa wind (vectors), jet stream speed (contour); (right column) mean annual precipitation (shading) and monsoon-like seasonality according to the MPI index (red dots). Sensitivity tests are compared to the LMio_smallT simulation (a, b, c) and anomalies (d to o) are expressed in order to highlight the impact of the uplift (i.e. LMio_smallT-test). We refer the reader to Figure 6 for an extended description of the panels.

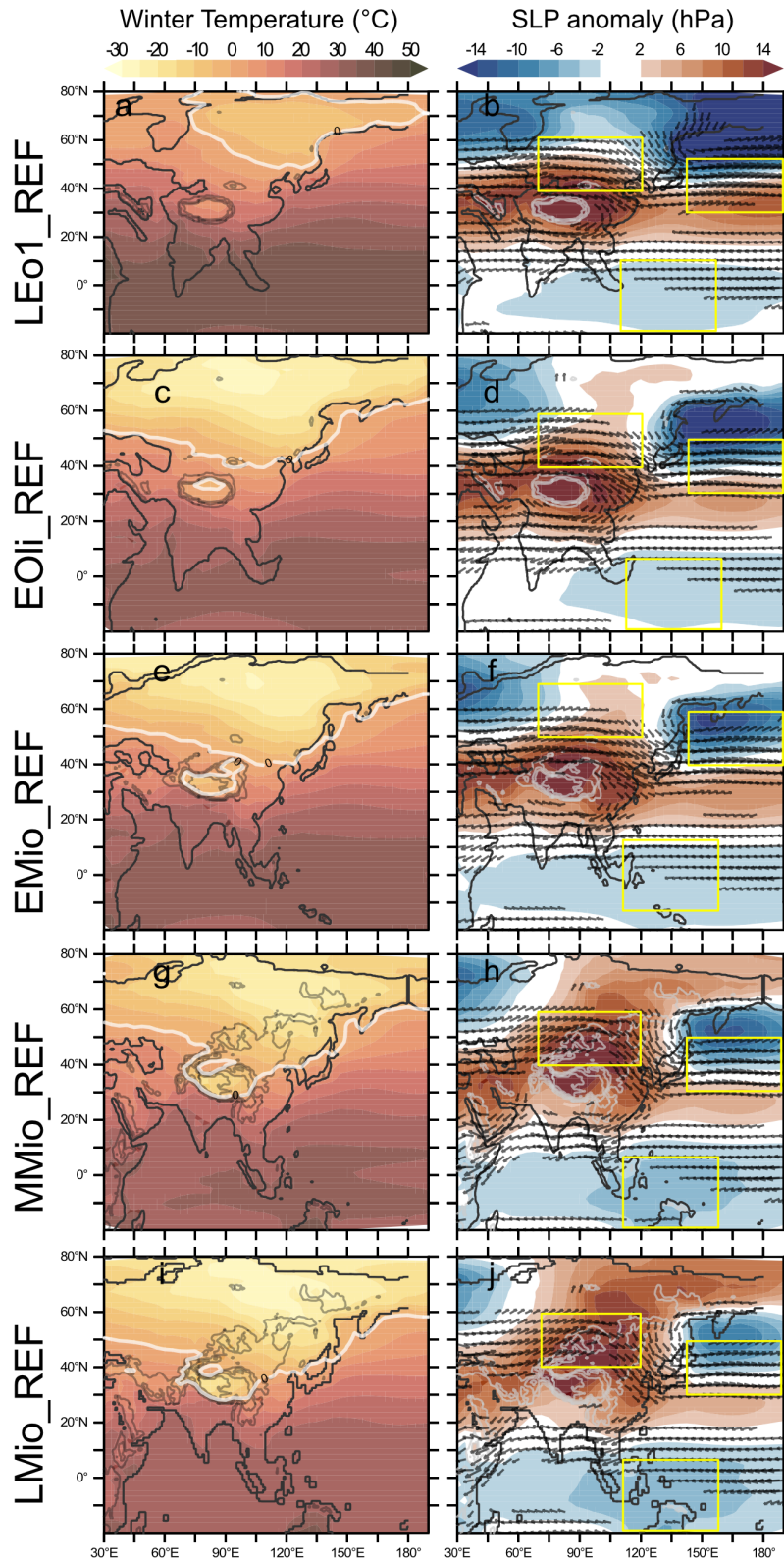


Figure 12. Evolution of the winter monsoon diagnostics throughout the Cenozoic. **Left column:** December-February (DJF) mean temperature at 2 m (shading, Celsius), and 0 °C isotherm (white contour); **Right column:** DJF normalized Sea Level Pressure anomaly (shading, hPa) and 850 hPa winds above 4 m/s (vectors). In all captions, topography is overlain in gray contour each 1000 m. Yellow boxes indicate the regions over which the sea level pressure gradients are calculated for the East Asian Winter Monsoon Index, presented in section 4.3

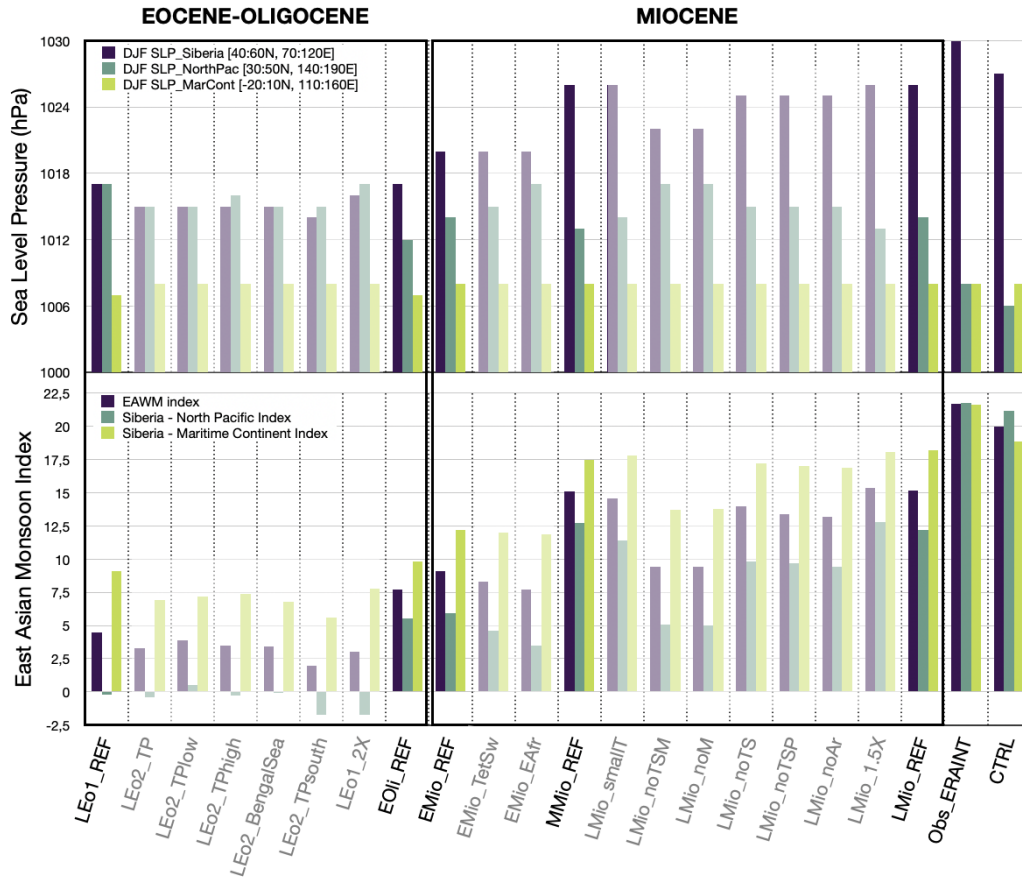


Figure 13. Upper panel: December-February (DJF) Sea Level Pressure (hPa) evolution over Siberia, North Pacific and the Maritime Continent ; lower panel: Evolution of the East Asian Winter Monsoon index (purple) and of the East-West and North-South regional indexes throughout the different reference and sensitivity experiments, after (L. Wang & Chen, 2014).

1124 6 Discussion

1125 A long-standing debate pits the effect of continentality against that of orography as
 1126 the main driver of Asian climate evolution (Prell & Kutzbach, 1992; Kutzbach et al., 1993;
 1127 Ramstein et al., 1997; Fluteau et al., 1999; Z. Zhang et al., 2007; Roe et al., 2016; Zoura
 1128 et al., 2019; R. Zhang et al., 2021). Our results (summarized in Table 2) suggest that
 1129 these two features impacted different climatic characteristics of the Asian monsoons and all
 1130 contributed to the emergence of current Asian climate. In the following sections, we put our
 1131 main findings in perspective with available paleoclimatic indicators and previous modeling
 1132 work, and tackle specific questions raised in section 2.4.

1133 6.1 Summer monsoons in the Eocene: driven by $p\text{CO}_2$, Tibetan paleo- 1134 geography or continentality ?

1135 Although the LEO1_REF experiment simulates the seasonally wet climate indicated
 1136 by eastern Chinese floras well (Meng et al., 2018; Y. Xie et al., 2022), it fails to identify the
 1137 subhumid to humid climate inferred from fossil paleoflora and paleofauna in northern India
 1138 (Saxena & Trivedi, 2009) and Tibet (T. Su et al., 2020; Sorrel et al., 2017). In a previous
 1139 study, this south Asian aridity bias in our experiment was interpreted as resulting from the
 1140 hot surface temperatures simulated in India, together with strong mid-tropospheric subsi-
 1141 dence that inhibits deep convection and condensation, especially over northern India (Tardif
 1142 et al., 2020). However, a previous study using the same model and paleogeography was able
 1143 to simulate monsoon-like climate in both the EAM and SAM domains, providing that or-
 1144 bital configurations induced summer solar maxima (Tardif et al., 2021). This orbitally-paced
 1145 wet/dry climate dichotomy is, at least qualitatively, in agreement with stratigraphic and pa-
 1146 leobotanic evidence from southern and eastern China (C. Huang & Hinnov, 2019; Licht et
 1147 al., 2014) and hints at a strong sensitivity of the South Asian region to surface tempera-
 1148 ture gradients in the Eocene. Such sensitivity seems further confirmed by the fact that, in
 1149 the present study, South Asian Eocene climate appears especially sensitive to experiments
 1150 affecting land-sea distribution and global temperature (LEO1_2X and LEO2_BengalSea, SI
 1151 Fig. 15 f, g).

1152 The East Asian monsoon domain, on the other hand, displays sensitivity to topo-
 1153 graphic changes in the Tibetan region. We simulate that an uplift of a proto-TP (LEO2_TP
 1154 - LEO2_TPLow) would result in an onset or increase in EAM monsoon-like precipitation
 1155 seasonality due to increased summer precipitation (up to 50%, Fig. 6 e, SI Fig. 15 b, c,
 1156 d), in agreement with previous modeling work (R. Zhang et al., 2018; R. Zhang, Jiang, &
 1157 Zhang, 2017; R. Zhang et al., 2021; Z. Zhang et al., 2007; X. Liu & Yin, 2002; X. Li et al.,
 1158 2017; Farnsworth et al., 2019; Thomson et al., 2021). Likewise, the decreased EAM intensity
 1159 and increased inland Asian aridity when the southern edge of the TP is situated at 30°N is
 1160 coherent with a previous study testing the impact of the latitudinal position of the incipient
 1161 TP (R. Zhang et al., 2018). Therefore, the inception or reinforcement of the East Asian
 1162 summer monsoon in the Eocene seems intertwined with the TP evolution history, as well
 1163 as with existing topography before the TP uplift (e.g. the Lhasa terrane), although their
 1164 paleoelevation and extent is not well constrained. Furthermore, we show that monsoon-like
 1165 rainfall seasonality may have occurred prior to the establishment of modern-like East Asian
 1166 winter monsoon circulation and Jet Stream summer migration, which were likely established
 1167 during the late Paleogene or the Neogene.

1168 6.2 Forcing factors of wintertime climate variability

1169 The observation of loess deposits in northeastern Tibet as early as 40 Ma was in-
 1170 terpreted as resulting from the periodic inception (or reinforcement) of the Siberian High
 1171 in winter due to enhanced continentality after the Paratethys Sea retreat (Meijer et al.,
 1172 2021). Other loess deposits reported during the Oligocene in the Chinese Loess Plateau
 1173 and Mongolia, apparently coeval with periods of cooling and ice-sheet expansion (J. Sun &

1174 Windley, 2015; Wasiljeff et al., 2022), further suggest the presence of a Siberian High and
 1175 corresponding dust transport in the late Paleogene. The potential forcing factors driving the
 1176 establishment of this winter pressure pattern have, however, been significantly less studied
 1177 than summer monsoons dynamics, with studies mostly focusing on the impact of orography
 1178 rather than continentality (White et al., 2017; Sha et al., 2015; Baldwin & Vecchi, 2016).
 1179 Our results suggest that both increased continentality and regional uplift contributed to-
 1180 gether to the progressive inception of the modern Siberian High, while $p\text{CO}_2$ decrease alone
 1181 had a lesser impact (Fig. 13). A modern-like Siberian High with typical EAWM wind
 1182 patterns is however not simulated until our middle and late Miocene simulations, when the
 1183 Tian Shan, but especially the Mongolian landforms, are set to their present-day elevation
 1184 in the model.

1185 Previous modeling work based on present-day geography that tested the effect of Mon-
 1186 golian landform uplift (Sha et al., 2015; Shi et al., 2015; Yu et al., 2018; R. Zhang, Jiang,
 1187 Zhang, Cheng, & Zhang, 2017) reproduced the same tendencies that we present here, namely
 1188 a strong increase of the winter sea level pressure over Siberia and typical deviation of the
 1189 westerlies towards northern latitudes resulting in a pronounced drying of eastern Asia and
 1190 EAWM amplification (Fig. 11 d). Other studies focusing more on the summer monsoon
 1191 simulate increased EASM precipitation and wind circulation, although they incorporate
 1192 Tian Shan uplift together with Mongolia (Yu et al., 2018; R. Zhang et al., 2012; H. Tang
 1193 et al., 2013; Zhisheng et al., 2001), which remains coherent with our findings testing the
 1194 uplift of both features (Fig. 11 h). Although the effect of Mongolian landforms has not
 1195 been explicitly tested in paleoclimate modeling studies using realistic paleogeographic re-
 1196 constructions, we can generally observe that when a modern-like Mongolian orography is
 1197 prescribed, the simulated winds and precipitation are close to the modern-like EAWM (Licht
 1198 et al., 2014; X. Li et al., 2017), while this is not the case with a reduced Mongolia eleva-
 1199 tion (Huber & Goldner, 2012). Given the large uncertainties related to the Hangay Dome
 1200 uplift history, performing additional experiments testing the impact of an early uplift in an
 1201 Oligocene configuration could enable one to evaluate the potential presence of a Siberian
 1202 High at that time, and therefore help interpret early loess-like deposits observed at that
 1203 time in the Chinese Loess Plateau and Mongolia (J. Sun & Windley, 2015; Wasiljeff et al.,
 1204 2022). Loess-like deposition in the Xining Basin at ~ 40 Ma, however, seems too old to
 1205 be attributed to the development of the modern Siberian High, given that Mongolia was
 1206 probably at low elevation and that continentality alone does not seem sufficient to trigger a
 1207 modern-like Siberian anticyclone. This suggests that alternative mechanisms for dust trans-
 1208 port should be explored for this period, although testing this hypothesis would require new
 1209 data such as daily to hourly wind speed model outputs (Aoki et al., 2005).

1210 **6.3 The critical role of peripheral landforms in the establishment of mon-** 1211 **soon seasonal features in the Miocene**

1212 In agreement with previous studies (R. Zhang, Jiang, & Zhang, 2017; Acosta & Huber,
 1213 2020; H. Tang et al., 2013), our results highlight the key role played by the Anatolian-
 1214 Iranian Plateau uplift in modern-like SAM establishment, as it reinforces both summer
 1215 rainfall and winter aridity (Fig. 9 j, k). The effect of the Tian Shan uplift on Asian climate
 1216 is less conspicuous, likely owing to the proximity of these mountains to other important
 1217 landforms (e.g. Pamir-Karakoram, Anatolian-Iranian Plateau, Mongolia), which may result
 1218 in competing effects on the regional climate. While a rigorous comparison between our
 1219 results and previous studies is hampered by many diverging boundary conditions, all tend
 1220 to indicate that the Tian Shan uplift would have: i) increased yearly orographic precipitation
 1221 on the ranges, ii) decreased precipitation in inland Asian mid-latitudes and iii) reinforced the
 1222 EAM through increased summer precipitation (Fig. 11 m, n, o) (Baldwin & Vecchi, 2016;
 1223 H. Tang et al., 2013; Yu et al., 2018; R. Zhang, Jiang, & Zhang, 2017; R. Zhang, Jiang,
 1224 Zhang, Cheng, & Zhang, 2017). The key role played by the Tian Shan ranges (together with
 1225 the Pamir and Mongolian landforms) in triggering the Jet Stream northward migration in
 1226 summer (SI Fig. 18 e, f, Fig. 9 e, h, k, n), as suggested by previous modeling work derived

1227 from present-day geography (Sha et al., 2020; Shi et al., 2015), likely contributes to this
1228 EASM intensification. Substantial differences are nevertheless noted with other modeling
1229 studies, especially regarding the impact of Tian Shan uplift on inland Asian and northeastern
1230 Indian climate, likely due to these regions' key location on the westerlies path. Westerlies
1231 may indeed be deviated either to the North or the South, depending on the respective
1232 degree of elevation of the Iranian-Anatolian Plateau and the Tian Shan. While we simulate
1233 a weakening of SAM rainfall seasonality due to increased winter precipitation after the Tian
1234 Shan uplift, the opposite trend is simulated in studies where the Anatolian-Iranian Plateau
1235 is already fully uplifted at the time of the Tian Shan uplift, thus preventing the westerlies
1236 from bringing moisture to India (Baldwin & Vecchi, 2016; H. Tang et al., 2013; Yu et al.,
1237 2018; R. Zhang, Jiang, & Zhang, 2017; X. Wang et al., 2020).

1238 Our set of simulations also highlights that East African topography can strongly impact
1239 the Neogene evolution of the SASM as it greatly contributes to the establishment of modern-
1240 like atmospheric circulation, as discussed by Sarr et al. (2022). The presence of intermediate
1241 to high elevation topography in East Africa in the early Miocene is for example crucial
1242 for the "early" bending of the Somali Jet, before the closure of the West Tethys Seaway
1243 and establishment of Anatolian-Iranian topography with modern-like elevation (Fig. 9 e).
1244 Increasing elevation in Eastern Africa is also responsible, in our simulations, for increasing
1245 SASM precipitation. This contrasts with previous studies that simulate reduced SASM
1246 precipitation, due to African topography blocking the advection of moisture from equatorial
1247 Africa towards India (H.-H. Wei & Bordoni, 2016; Chakraborty et al., 2009). We hypothesize
1248 that this difference may come from these studies using present-day paleogeography, and/or
1249 unchanged fixed sea surface temperatures for all the experiments, thereby neglecting the
1250 potential effect of the East African uplift on Indian Ocean surface temperature gradients.
1251 It is also very likely that the impact of East African highlands uplift on South Asian rainfall
1252 regime may be modulated by the degree of uplift in the Iranian-Anatolian region. Better
1253 constraints on the relative degree of uplift of these different mountain ranges in the past,
1254 although seldom taken into account, is therefore critical to interpreting paleoenvironmental
1255 indicators.

1256 **6.4 Continentality increase in the Miocene and acts as an amplifier of** 1257 **SAM, EAM and inland aridity**

1258 The northward motion of the ITCZ over the Arabian Sea and the stronger SASM
1259 precipitation in response to continentality increase in the Miocene is a feature clearly sim-
1260 ulated in our experiments, and which is obtained with either the Paratethys Sea retreat or
1261 an emerging Arabian Peninsula (Fig. 8 e, h). The role of the Paratethys Sea retreat as a
1262 vector of amplified summer low pressure and resulting moisture advection via a strength-
1263 ened Somali Jet is well established by modeling experiments (Ramstein et al., 1997; Fluteau
1264 et al., 1999; R. Zhang et al., 2021; Z. Zhang et al., 2007). The Paratethys Sea retreat
1265 also drastically dries the winter-time "westerly corridor" (centered around 40°N, Fig. 8 h),
1266 resulting in increased inland Asian aridity (Fig. 8 i), here again mostly in coherence with
1267 previous modeling studies (R. Zhang et al., 2021; Z. Zhang et al., 2007), and field evidence
1268 (Palcu et al., 2021). On the other hand, the effect of increased land exposure in the Arabian
1269 Peninsula region on Asian climate has been, to our knowledge, significantly less explored by
1270 modeling studies and is rarely mentioned as a potential forcing factor in proxy data studies.
1271 We tentatively attribute this to the lack of constraints and data in this region. Given the
1272 shallow depth and extension of the Arabian Peninsula (roughly comparable to India's area),
1273 it is very likely that the episodes of partial or complete flooding that occurred throughout
1274 the Oligocene to late Miocene (Barrier et al., 2018; Otero & Gayet, 2001) may have had
1275 great consequences on atmospheric circulation patterns. Following Sarr et al. (2022), our
1276 results show that increased peninsula exposure reinforces the Somali Jet in summer and en-
1277 hances summer precipitation in SAM and EAM domains (Fig. 8 e). Additionally, it drives
1278 the onset of upwelling in the Arabian Sea (see Sarr et al. (2022)), which is traditionally
1279 used as a proxy for the establishment of a monsoon wind regime in South Asia. Our simu-

1280 lations suggest that a proto Somali Jet and intense summer precipitation were most likely
1281 established in the middle Miocene, before the Arabian Peninsula emergence, and therefore
1282 before any strong upwelling activity initiated in the Arabian Sea. In addition, the onset of
1283 upwelling in the Arabian Sea was likely uncorrelated to potential Miocene uplift events in
1284 the Himalaya-Tibet region (Sarr et al., 2022).

1285 **6.5 Persistent uncertainties and directions for future work**

1286 Our study demonstrates the key role played by land-sea distribution and by all to-
1287 pographic features surrounding the Indian Ocean in shaping the modern SAM, EAM and
1288 inland Asia aridity. Nevertheless, additional work is required both from modeling and in the
1289 field to help refine our understanding of the mechanisms at play throughout the Cenozoic
1290 in Asia. By showing the influence of several geographical units on Asian climate during
1291 different geological periods, our results highlight the conceptual limitations of performing
1292 tests based on present-day geographies or assuming homogeneous Asian orographic uplift
1293 throughout geological history to evaluate past monsoon evolution. In that respect, it is not
1294 surprising that most of those type of studies simulate modern-like summer and winter mon-
1295 soons features when maintaining high-elevation African, Mongolian and Anatolian-Iranian
1296 landforms, as well as an emerged Arabian platform (Roe et al., 2016; Zoura et al., 2019;
1297 Acosta & Huber, 2020). Our results also suggest that some paleogeographic events consid-
1298 ered roughly coeval may have had opposite influences on the Asian climate’s seasonal fea-
1299 tures, for example the Tian Shan and Anatolian-Iranian Plateau uplift discussed in section
1300 6.2. Refining the relative chronology of these various uplift phases is therefore paramount to
1301 better interpret the variations in monsoon intensity recorded by proxy data. Additionally,
1302 paleoelevation thresholds are likely to be critical for constraining Asian climate evolution,
1303 due to interactions of the topography with either low level (e.g. the summer monsoon winds)
1304 or mid tropospheric (e.g. the Jet Stream) air masses (X. Liu & Yin, 2002; X. Wang et al.,
1305 2020; Zhisheng et al., 2001; Prell & Kutzbach, 1992).

1306 We also acknowledge that a better understanding of the seasonal monsoon mechanisms
1307 simulated here would benefit from additional simulations performed with increased spatial
1308 and temporal resolution. Previous work have for example demonstrated the importance of
1309 high spatial resolution in the full depiction of the SASM circulation (Acosta & Huber, 2017)
1310 or the study of the interplay between the summer Jet Stream behavior and the formation of
1311 the Meiyu-Bayu front in Eastern Asia (Kong et al., 2017; Sampe & Xie, 2010). Additional
1312 diagnostics at planetary-scale, may also provide interesting comparison with previous studies
1313 on the impact of monsoon dynamics on large scale circulation patterns and long-distance
1314 teleconnections (Rodwell & Hoskins, 1996; Merlis et al., 2013; Biasutti et al., 2018).

1315 A more precise view of land-sea distribution and regional seaway evolution is also
1316 paramount. Indeed, these factors influence SST gradients, thereby controlling sea level
1317 pressure patterns, low tropospheric circulation, and oceanic circulation. Although our re-
1318 sults point to a minimal effect of the mechanical closure of the Tethyan Seaway on regional
1319 climate, other features need to be tested. For example, a better understanding of the In-
1320 donesian Gateway and Indonesian throughflow evolution would be of great interest, as these
1321 control in part Indo-Pacific Warm Pool SST distribution (Cane & Molnar, 2001; Tan et
1322 al., 2022) and may thus have influenced Asian climate (Q. Su et al., 2019). While the final
1323 step of the Indonesian Gateway constriction likely occurred during the Pliocene (Auer et
1324 al., 2019), its paleobathymetry and complex geometric configuration remain largely undoc-
1325 umented between 17-3.5 Ma (Kuhnt et al., 2004). While no major tectonic deformation is
1326 thought to have occurred during this period, the shallow depth of the gateway likely made it
1327 sensitive to middle Miocene sea level fluctuations (Sosdian & Lear, 2020), with possible im-
1328 pact on the Eastern Indian and Western Pacific Oceans surface temperatures and therefore
1329 on SST gradients. Although more distant from Asia, studies have also pointed to a possible
1330 influence of the Central American Seaway (Prabhat et al., 2022) or Drake (Toumoulin et

1331 al., 2020) and Tasman (Scher et al., 2015) Gateway openings on oceanic circulation, which
1332 may have in turn impacted sea surface temperature gradients and monsoon activity.

1333 7 Conclusions

1334 In this study, we took advantage of recent work that extended the Asian paleo-
1335 monsoonal record back to the Paleogene and of the improved understanding of the physical
1336 concepts underpinning the monsoons derived from climate model sensitivity experiments.
1337 Overall, we simulate an intensification of monsoonal metrics throughout the Cenozoic, in
1338 response to evolving paleogeography and global climate trends. Our results suggest that
1339 modern-like monsoons likely existed since at least the mid-late Miocene when considering to-
1340 gether patterns of atmospheric circulation as well as seasonality and precipitation amounts,
1341 in agreement with paleoclimate indicators. However, we also show that high rainfall sea-
1342 sonality already existed since the Paleogene in both SAM and EAM domains, before the
1343 establishment of modern atmospheric circulation patterns. In accord with previous studies,
1344 our results highlight the limited effect of $p\text{CO}_2$ on monsoon dynamics, when compared to
1345 the effect of paleogeographic changes. A notable exception is the strong impact of a $p\text{CO}_2$
1346 drawdown at the EOT on the northward migration of the subtropical Jet Stream during
1347 summer, which increases precipitation in the EAM region.

1348 Our results also strongly suggest that any variations in the land-sea distribution,
1349 whether due to tectonic activity or sea level fluctuations, would have strongly impacted
1350 Asian climate, despite the fact that the exact shape and extent of coastal shallow-water
1351 environments remain poorly constrained. Many episodes of increased continentality are doc-
1352 umented over the course of the Cenozoic, including regional regressions of the Paratethys
1353 epicontinental Sea and periodic exposure of the Arabian Platform due to sea level fluctua-
1354 tions since the Eocene-Oligocene Transition. All of these events likely favored the enhance-
1355 ment of summer moisture penetration into Asia, by promoting the extension of the wide
1356 subtropical low pressure belt, thereby amplifying the seasonal latitudinal migration of the
1357 ITCZ on land.

1358 In addition, while the evolution of the Himalayan-TP complex is recurrently put for-
1359 ward as the leading phenomenon driving Asian climate evolution, our results rather point
1360 towards a combined effect of all landforms surrounding the modern Indian Ocean. We also
1361 emphasize that all paleogeographic features do not bear the same importance on the SASM,
1362 EASM or EAWM, respectively. Our simulations suggest that the Tibetan Plateau uplift
1363 likely favored the onset of monsoon-like precipitation seasonality in eastern Asia in the
1364 Eocene, by promoting the advection of moist air masses in summer and partially shielding
1365 the region from westerly moisture input during winter. Alongside that, we show that the up-
1366 lift of Eastern African and Anatolian-Iranian landforms was critical to intensify and redirect
1367 the Somali Jet towards the Indian subcontinent, thereby contributing to the development of
1368 a modern-like South Asian monsoon summer circulation and amplifying moisture advection
1369 towards Asia in summer. In addition, in our simulations, the uplift of the Anatolian-Iranian
1370 orogen contributed to enhancing winter aridity in most of southeastern Asia by blocking
1371 moisture advection from the westerlies to lower latitudes.

1372 Our simulations further show that uplift of the Tian Shan and the Mongolian moun-
1373 tains was key for the establishment of the winter Siberian High, and resulting aridification
1374 of inland and eastern Asia during winter. The summer Jet Stream migration north of the
1375 Tibetan Plateau was likely amplified in the mid-late Miocene, as the Pamir and Tian Shan
1376 landforms approached their modern elevations and contributed to channeling the jet towards
1377 higher latitudes. This relocation of the Jet Stream amplifies the penetration of moist air
1378 masses over eastern China, which contributes to increasing EAM precipitation seasonality.
1379 Overall, our results do suggest a fundamentally polygenetic history of monsoonal evolution,
1380 in agreement with available paleoclimate indicators and previous modeling studies. This
1381 suggests that the very concept of a "monsoon onset" is inappropriate. We also stress the

1382 need for more accurate constraints on the evolution of paleo-shorelines, the relative timing
1383 of peripheral mountain belts uplift, as well as on the paleo-elevation of landforms in order
1384 to better understand the evolution of Asian climate in the past.

1385 **Acronyms**

1386 **ITCZ** Inter Tropical Convergence Zone
1387 **SASM** South Asian Summer Monsoon
1388 **EASM** East Asian Summer Monsoon
1389 **EAWM** East Asian Winter Monsoon
1390 **TP** Tibetan Plateau
1391 **EOT** Eocene Oligocene Transition
1392 **CLAMP** Climate Leaf Analysis Multivariate Program

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1409 processing/IPSL_Boundary_Conditions/](https://paleoclim-cnrs.github.io/documentation-processing/IPSL_Boundary_Conditions/)). The scientific color maps (<https://www.fabiocrameri.ch/colourmaps/>)
1410 (Crameri, 2021) used in most of the Figures prevent visual distortion of the data and exclu-
1411 sion of readers with color vision deficiencies (Crameri et al., 2020).

1412

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3126 8 Supplementary

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55	(Beasley et al., 2021)
56	(Zhuang et al., 2017)

Table 3. List of references used in Figure 2 for paleoclimate indicators compilation

	Reference (suite)
57	(Page et al., 2019)
58	(Q. Yuan et al., 2021)
59	(Meng et al., 2018)
60	(Miao et al., 2013)
61	(Z. Zhang & Sun, 2011)
62	(J. Sun & Zhang, 2008)
63	(Q.-g. Sun et al., 2002)
64	(J. Sun et al., 2020)
65	(T. Su, Spicer, et al., 2019)
66	(M. Wu et al., 2021)
67	(Song et al., 2020)
68	(Miao, Wu, et al., 2016)
69	(D. Rea et al., 1985)
70	(Pettke et al., 2002)
71	(Ziegler et al., 2007)
72	(B. Li et al., 2016)
73	(Garziona et al., 2005)
74	(Qiang et al., 2011)
75	(Bougeois et al., 2018)
76	(W. Yang et al., 2019)
77	(J. Lee et al., 2020)
78	(Bolton et al., 2022)
79	(Hui et al., 2021)
80	(Miao, Song, et al., 2016)
81	(H. Zhao et al., 2020)
82	(C. Yang et al., 2021)
83	(L. Zhao et al., 2020)
84	(Heermance et al., 2018)
85	(J. Liu et al., 2016)
86	(Jiang & Ding, 2008)
87	(Peng et al., 2016)
88	(Suarez et al., 2011)
89	(Barbolini et al., 2020)
90	(Y. Ma et al., 2005)
91	(H. Wang et al., 2019)
92	(Jiang et al., 2017)
93	(W. Liu et al., 2014)
94	(Kaakinen et al., 2006)
95	(Y. Wang & Deng, 2005)
96	(F. Li et al., 2008)
97	(Z. Liu et al., 2014)
98	(Miao et al., 2011)
99	(Charreau et al., 2012)
100	(Holbourn et al., 2018)
101	(G. Wei et al., 2006)
102	(Y. Huang et al., 2007)
103	(Gupta et al., 2015)
104	(X. Yang et al., 2020)
105	(Bialik et al., 2020)
106	(H. Tang et al., 2020)
107	(H. Huang et al., 2021)
108	(Hoorn et al., 2000)
109	(Polissar et al., 2021)
110	(Y. Xie et al., 2022)

Table 4. (suite) List of references used in Figure 2 for paleoclimate indicators compilation

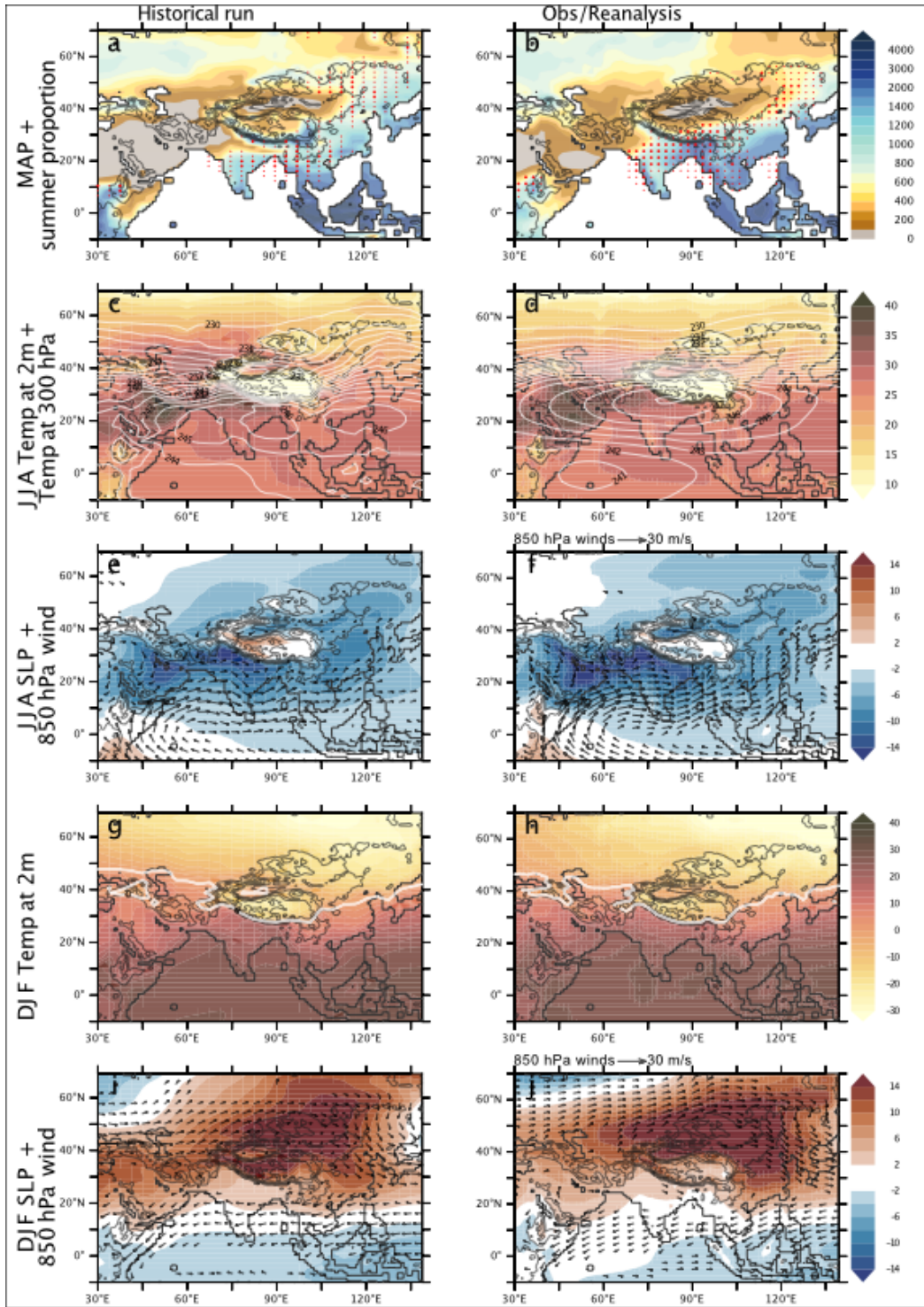


Figure 14. Comparison of monsoon diagnostics obtained in our control historical simulation (**Left column**) with modern observations and reanalysis (**Right column**). (**a, b**) Mean Annual Precipitations (shading, mm/yr), and regions where the Monsoon Precipitation Index is over 0.5 (thin red dots) and 0.75 (thick red dots); (**c, d**) JJA mean 2 m temperature (shading, Celsius) and 300 hPa atmospheric temperature (white contour, Kelvin); (**e, f**) JJA normalized Sea Level Pressure anomalies (shading, hPa) and 850 hPa winds (vectors over 4 m/s); (**g, h**) DJF mean 2 m temperature (shading, Celsius); (**i, j**) DJF normalized Sea Level Pressure anomalies (shading, hPa) and 850 hPa winds (vectors over 4 m/s). In all captions, topography in gray contour, each 1000 m. On the right column, precipitation are from GPCP, pressure and 300hPa temperature from ERAINT, 850hPa winds and surface temperature from ERA40.

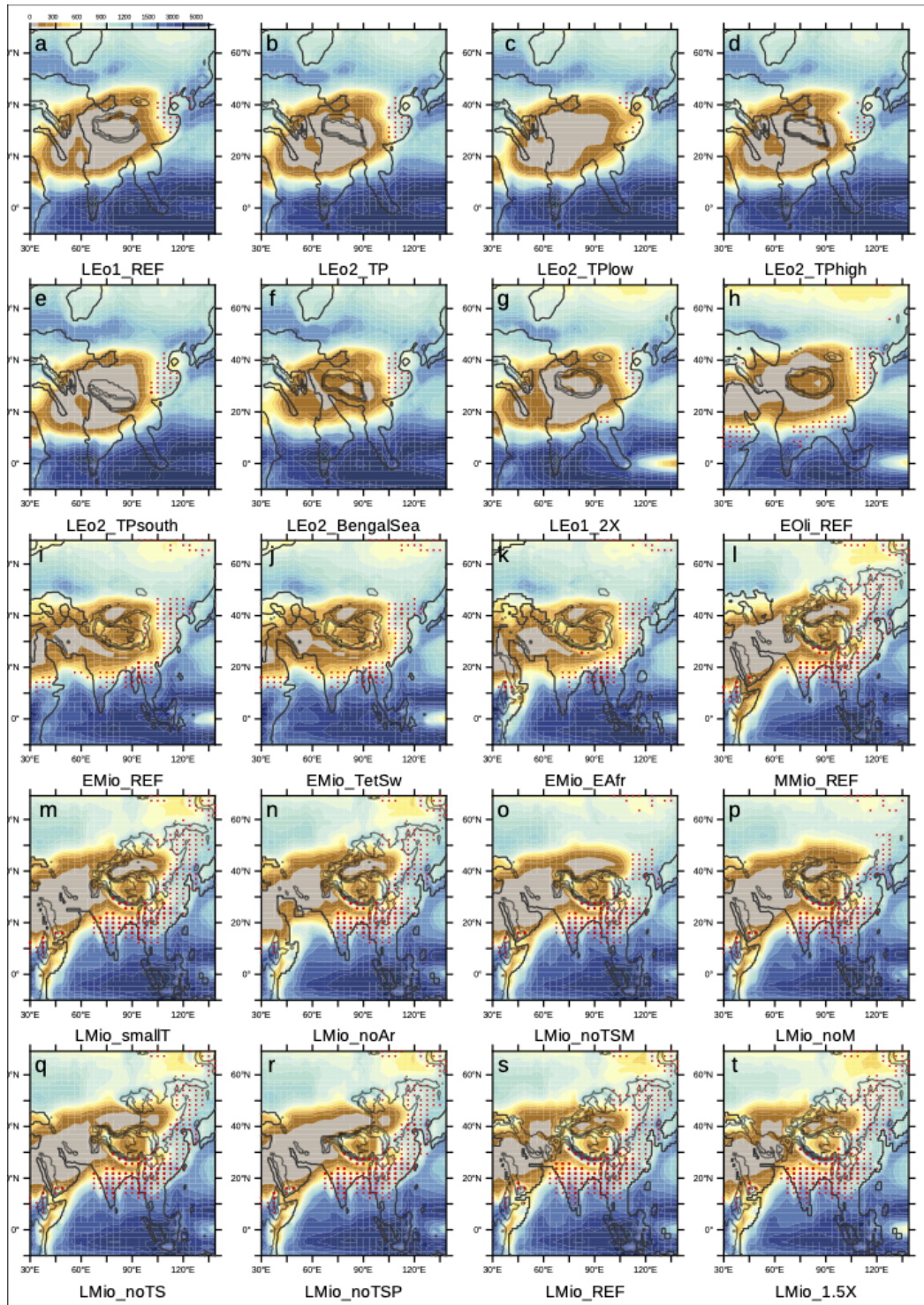


Figure 15. Mean Annual Precipitation (shading, mm/yr), overlain with regions where the Monsoon Precipitation Index (B. Wang & Ding, 2008) is over 0.5 (thin red dots) and 0.75 (thick red dots), for all reference and sensitivity experiments.

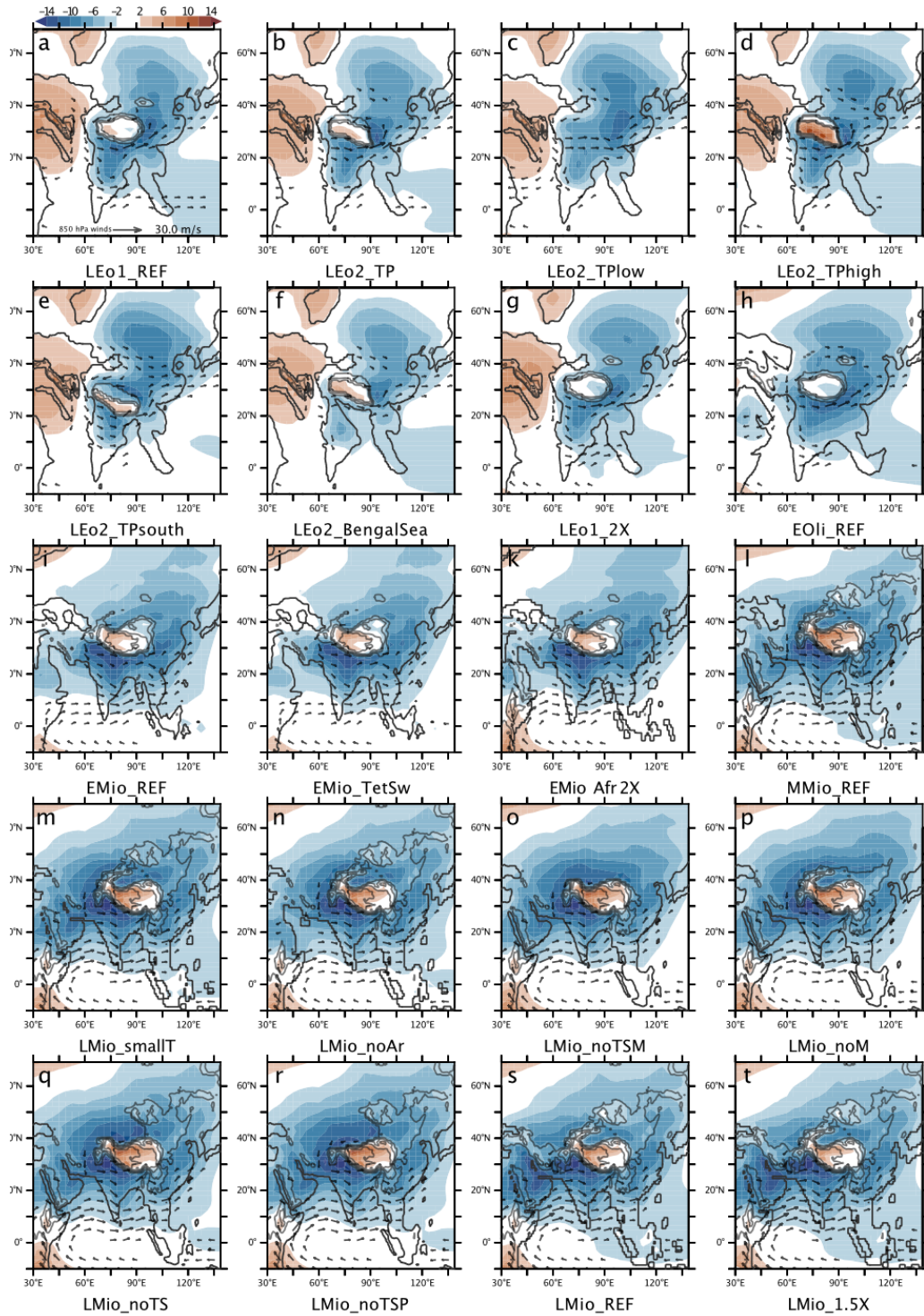


Figure 16. June-August (JJA) normalized Sea Level Pressure anomalies (shading, hPa) and 850 hPa winds over 4 m/s (vectors), for all reference and sensitivity experiments.

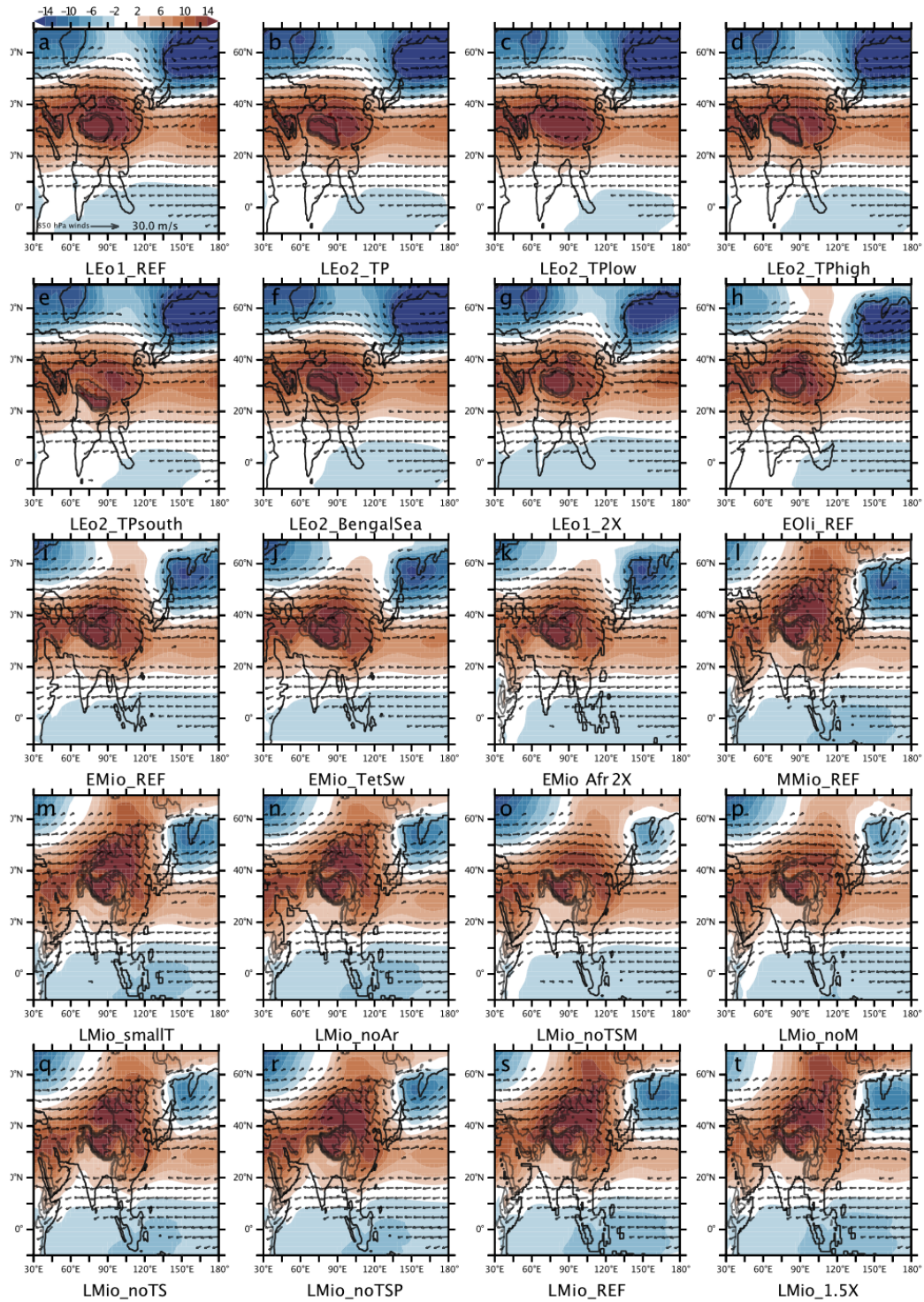


Figure 17. December-February (DJF) normalized Sea Level Pressure anomalies (shading, hPa) and 850 hPa winds over 4 m/s (vectors), for all reference and sensitivity experiments.

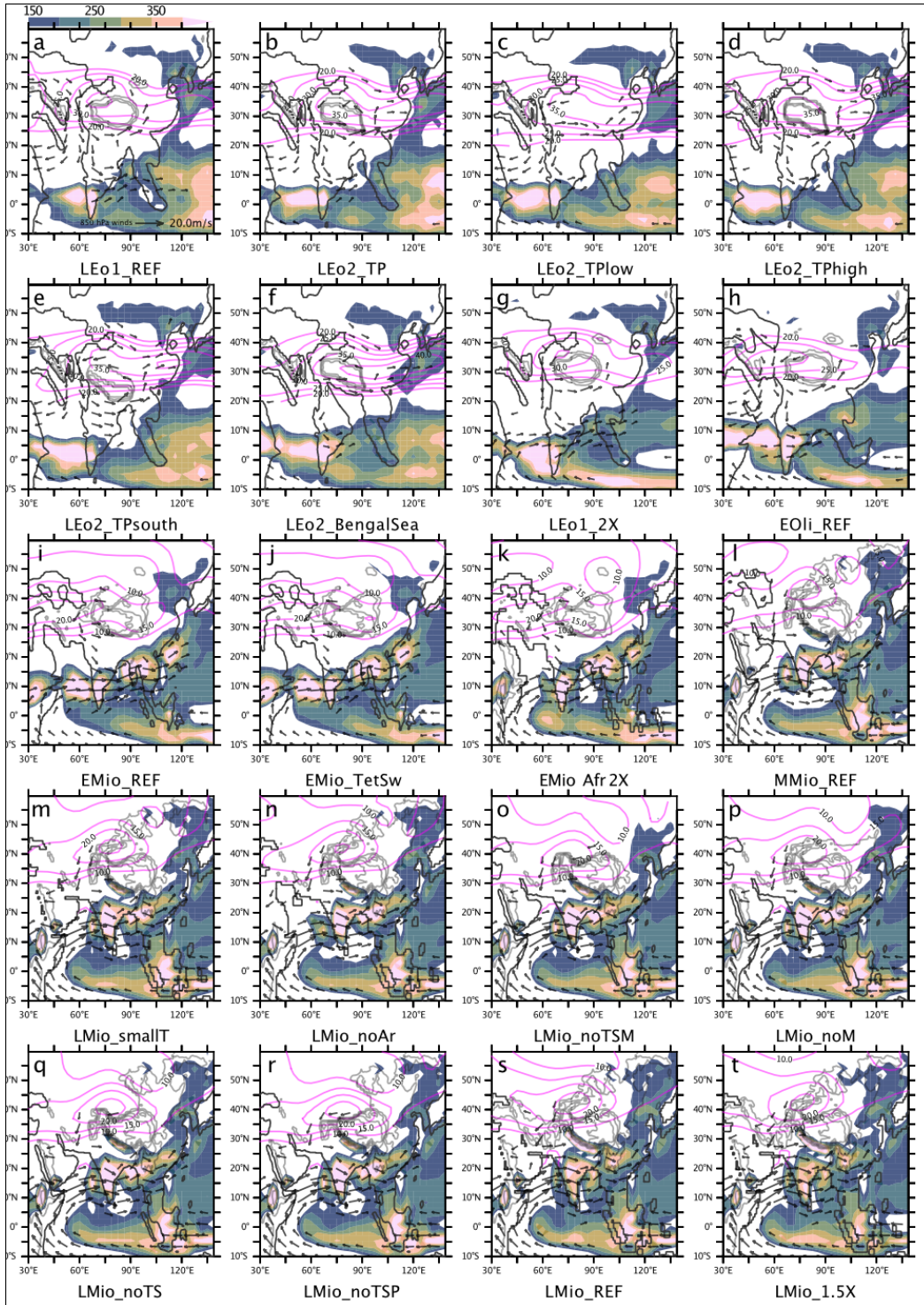


Figure 18. Jet Stream velocity and localization over Asia in August (purple outline), for all sensitivity experiments and precipitation for the month of August (shading, mm/month).