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

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# Asynchronous apparition of the Asian monsoons features since the late Eocene and associated paleogeographic and climatic drivers

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## Key Points:

- Asian monsoons modern characteristics appeared sequentially during the Cenozoic, driven by distinct paleogeographic and climatic forcings.
- Model results depict seasonal precipitation in parts of Southern and Eastern Asia since the Paleogene.
- The Paratethys Sea retreat and the Arabian Platform emergence, by greatly increasing continentality, are instrumental for the development of surface temperature gradients between Asia and the Indian Ocean which reinforces the advection of moist air towards Asia in summer.
- Iran-Zagros and East African landforms contribute to channel and enhance the Somali Jet that brings moisture to Southeastern Asia in summer, while the Himalayan-Tibetan plateau uplift induces heavy orographic precipitation.
- Mongolian landforms uplift is the main driver for the Siberian high pressure settlement in winter. It results in a broad reinforcement of East Asian Monsoon precipitation seasonality, as it strongly reduces winter precipitation. It also induces aridification of current Gobi desert region.
- Tian Shan and Pamir uplifts, by intercepting and deviating westerly moisture flux, concur to force the summer northern Jet Stream migration in Tibet and Eastern China. This triggers local orographic precipitation, increased winter precipitation in India (weaker South Asian Monsoon), and increased summer precipitation in North-eastern China (stronger East Asian monsoon). It enhances global aridification of Gobi, Taklamakan and Northeastern Tibet regions.

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## Abstract

Modern Asian monsoons are triggered by complex interactions between the atmosphere, Asian and African orography and the surrounding oceans, resulting in a highly seasonal climate characterized by typical regional features. Although long considered as a Neogene phenomenon, recent evidence for monsoon-like seasonality as early as in the Paleogene Greenhouse, has come to challenge this paradigm. The occurrence of monsoons in a climatic and paleogeographic context very different from the present-day question our understanding of the drivers underpinning this peculiar atmospheric phenomenon, in particular with regards to its dependence to paleogeography. In this study, we take advantage of the wealth of new studies to tentatively draw an up-to-date portrait of the Asian tectonic and paleoenvironmental evolution through the Cenozoic and of the questions that remain to be answered. We then simulate the evolution of the Asian monsoon subsystems by investigating seasonal trends of 20 paleoclimate simulations spanning the late Eocene to latest Miocene (40-6 Ma). At odds with the traditional view of a monsoonal evolution driven mainly by the Himalayan-Tibetan uplift, this study confirms the findings of previous works stating the importance of peripheral landforms on the different Asian monsoons subsystems. Eastern African and Anatolian-Iranian topography, as well as the Arabian Peninsula emergence all concur to shape the South Asian summer Monsoon. Additionally, we suggest that continentality, together with Mongolia, Tian Shan and Pamir orographic evolution, and  $p\text{CO}_2$  decrease, played important parts in the settlement of the East Asian Monsoon and in the aridification of inland Asia.

## 1 Introduction

The South and East Asian monsoons are highly seasonal climatic phenomena, that today conditions the livelihood of billions of people. During summer, moisture-loaded winds blowing from the Indian Ocean (Fig. 1) trigger important nutrients-rich upwellings and lead to heavy precipitation over Southeastern Asia, allowing for widespread agriculture and fisheries. In winter, the reversal in wind direction induces advection of cold, dry and dust-loaded air masses from Siberia over Eastern and Southern Asia (Fig. 1). The spectacular extension and intensity of the Asian monsoons, combined to the highly dynamic tectonic context of the region, have triggered a lot of interest from the scientific community, in order to understand potential links between orography, land-sea distribution, global climate and the monsoon.

While monsoons have long been viewed as an atmospheric phenomenon intimately intertwined with the Himalayan-Tibetan plateau uplift history (Molnar et al., 1993; Zhisheng et al., 2001; Tada et al., 2016; X. Liu, Guo, et al., 2015) and rooted in the Neogene, first with records of strong upwellings initiating in the Arabian Sea at  $\sim 8$  Ma (Kroon et al., 1991), then with evidence for massive dust deposition in the Chinese Loess Plateau since  $\sim 22$  Ma (Guo et al., 2002), a wealth of recent studies has come to challenge this paradigm. Indeed, aeolian dust deposits in the Xining basin (northeastern China, see Fig. 1 for localities mentioned in the text), as well as paleoflora and isotopic measurement in fossils indicating a seasonal climate in Myanmar (Licht et al., 2014) and in the Yunnan region (South China) (Sorrel et al., 2017; Fang et al., 2021; Zheng et al., 2022) have been interpreted as pieces of evidence for an active monsoon as early as in the late Eocene. Proposing a monsoonal onset in the Paleogene, in a warmer, ice-free climate, during the early stage of the Tibetan plateau uplift and with much lower continentality due to higher sea level and to the presence of the Paratethys Sea in the west (see the late Eocene paleogeography in Fig. 3a), radically questions our understanding of the forcings of the monsoons.

Climate modeling provides a unique opportunity to understand the part played by each potential forcing on the resulting climate. Overall, such studies, and independently of the model used (complexity, resolution) or the simulation protocol (i.e., derived from modern geography or with realistic paleogeography), have tended to highlight that paleogeography

87 out-compete global  $p\text{CO}_2$  variations or far-field ice-sheets effects when it comes to the SAM  
88 and EAM settlement and intensification over multi-million years time scales (Farnsworth  
89 et al., 2019; J.-Y. Lee et al., 2015; Zoura et al., 2019; Roe et al., 2016; Thomson et al.,  
90 2021). In a second time, sensitivity experiments of increasing complexity and precision,  
91 have helped disentangle the effect of the Himalaya-Tibet (and its sub units) uplift from  
92 that of peripheral landforms uplifts, such as the Tian Shan ranges, the Iranian-Anatolian  
93 orogen, eastern African domes, Mongolia landforms (H. Tang et al., 2013; Acosta & Huber,  
94 2020; R. Zhang et al., 2017), and from the effect of land-sea distribution changes (Fluteau  
95 et al., 1999; Z. Zhang et al., 2007). While modeling studies provide mechanistic basis to  
96 apprehend potential leverages for monsoon and aridity modulation, it remains paramount to  
97 dispose of robust constraints on the paleogeography and paleoclimate evolution, and on the  
98 relative chronology of these events. The past decades have been rich in both modeling and  
99 field studies, for example with marine cores that were extended back to the late Paleogene  
100 (W. Ding et al., 2021; S. Ali et al., 2021; Betzler et al., 2016; Beasley et al., 2021), and  
101 increased coverage in well-dated paleoclimate indicators, which offer new constraints on  
102 Cenozoic tectonic and paleoclimate.

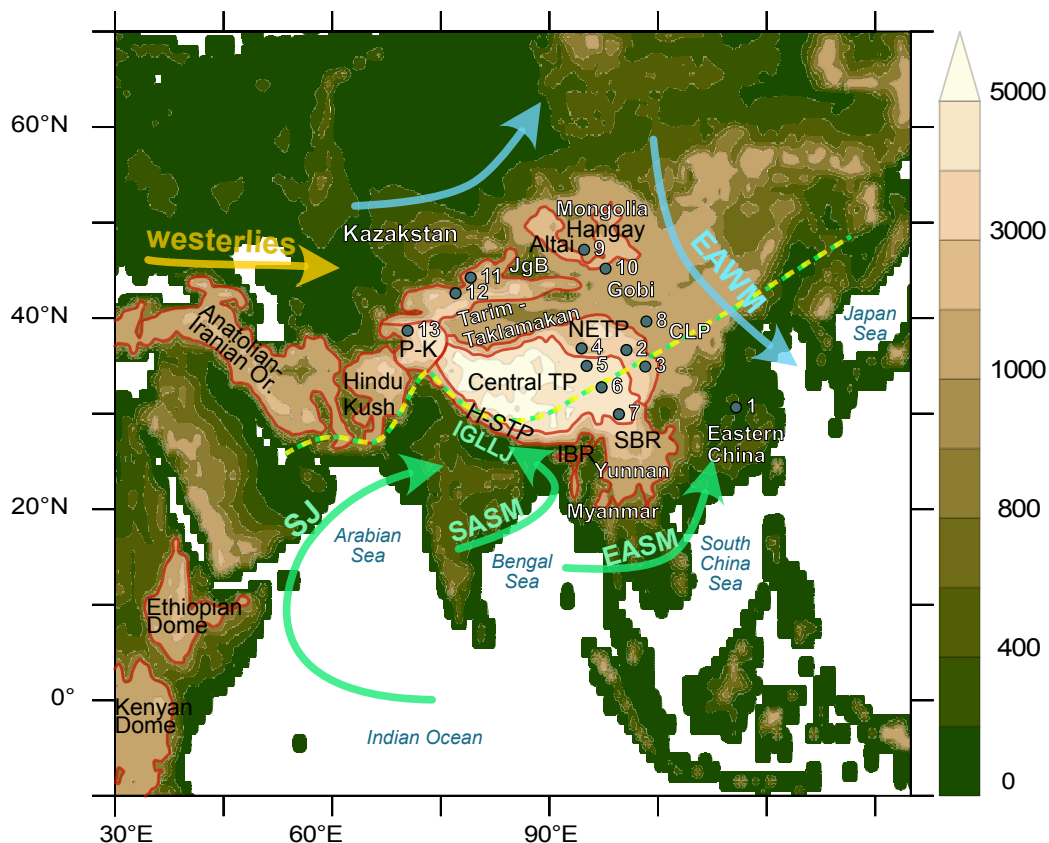
103 In this contribution, and building upon previous work (Sarr et al., 2022; Tardif et al.,  
104 2020; Barbolini et al., 2020), we aim at providing a comprehensive picture of large-scale wind  
105 circulation, precipitation seasonality and aridity evolution in continental Asia from the late  
106 Eocene to the late Miocene, using realistic paleoclimate simulations. In the following parts,  
107 we first describe in more depth the Asian monsoon subsystems characteristics (section 2),  
108 and propose an up-to-date overview of the Asian monsoon history as understood through  
109 proxy and models studies, and of remaining outstanding issues (section 3, Fig. 2). The main  
110 paleogeographic events that affected the regions surrounding the Indian Ocean together  
111 with continental Asia during the Cenozoic are presented (section 4), along with the past  
112 continental configurations that have been used (Fig.3). Other boundary conditions and  
113 simulations spin-up are then introduced, as well as criteria retained to track the Asian  
114 monsoons evolution throughout the different simulations (section 5). Our results are then  
115 presented and discussed in the light of actual knowledge.

## 116 2 Modern Asian Monsoon subsystems characteristics

117 The Asian monsoon is subdivided into two regional subsystems: the South Asian  
118 Monsoon, spreading roughly from Pakistan to Myanmar, and the East Asian Monsoon, from  
119 Myanmar to Japan. We hereafter refer to them as SAM and EAM respectively, SASM and  
120 EASM are used for the summer monsoon season while SAWM and EAWM refer to the winter  
121 monsoon season. These regional monsoons present specific climatic characteristics and have  
122 long been studied by distinct communities, although regions situated in the Southeastern  
123 Asian Peninsula can present mixed attributes of both subsystems and may be referred to  
124 as "transition area" in the literature (R. Spicer et al., 2017).

### 125 2.1 Summer season

126 The SASM is associated with intense orographic precipitation over the Himalayan  
127 foothills as well as over smaller mountains ranges like the Ghats in western India and the  
128 Indo-Burman Ranges in Myanmar (Fig. SI 14 a,b). In these regions, summer precipitation  
129 account for nearly 80% of the total annual precipitation and are responsible for massive fresh-  
130 water discharge in the Bay of Bengal, *via* the Ganga, Brahmaputra and Ayeyarwady Rivers  
131 and in the Arabian Sea *via* the Indus River. Precipitation is supplied by moisture-loaded  
132 air masses from the Inter Tropical Convergence Zone (ITCZ), which is steered towards the  
133 Asian continent during summer, as a consequence of the wide low pressure belt develop-  
134 ing from Arabia to eastern China (Fig. SI 14 d). This typical pressure pattern is driven  
135 by the establishment of a surface temperature gradient between the Indian Ocean and the  
136 overheated Asian continent and the Arabian Peninsula (up to 40°C, Fig. SI 14 d).



**Figure 1.** Modern geography and topography (in m), with main seasonal winds: Westerlies (yellow arrow), East Asian Winter monsoon (blue arrow) and Summer monsoons (green arrows); Somali Jet (SJ), Indo-Gangetic-Low Level Jet (IGLLJ). Dotted green/yellow line indicate the modern limit between the westerlies and the monsoonal regions. Main landforms units (in black) and main geographical regions (in white) mentioned in the text, with the following abbreviations: H-STP - Himalayas and Southern Tibetan Plateau, NETP - Northeastern Tibetan Plateau, IBR - Indo Burman Ranges, SBR - Sino Burman Ranges, ML - Mongolian Landforms, CLP - Chinese Loess Plateau, JgB - Junggar Basin. Smaller basins and localities mentioned in the text are numbered as follow: (1) Jiangnan B., (2) Xining B., (3) Linzhou B., (4) Qaidam, (5) Hoh Xil B., (6) Nangqian B., (7) Markam B., (8) Bayanhot B., (9) Valley of Lakes, (10) Taatsin Gol, (11) Ili B., (12) Issyk Kul B., (13) Tajik B.

137 Characteristic low-level wind patterns are associated with the SASM, represented by  
 138 the Findlater Jet (Findlater, 1969) (or Somali Jet (H.-H. Wei & Bordoni, 2016)), which  
 139 crosses the equator alongshore East Africa and blows towards the Indian subcontinent (Fig.  
 140 SI 14 d). This powerful trans-equatorial jet triggers strong surface oceanic currents, flowing  
 141 up the East African and Arabian coasts, then circumventing India from West to East. These  
 142 currents generate strong upwellings in the Arabian Sea and the Bay of Bengal (Schott & Mc-  
 143 Creary, 2001), favoring broad phytoplankton productivity blooms from July to September  
 144 (Lévy et al., 2007). Over the Bay of Bengal, the Somali Jet splits into two distinct branches  
 145 (Fig. SI 14 d), one pursuing its northeastern flow towards southeastern and eastern Asia  
 146 (from Myanmar to China), and the Indo-Gangetic Low Level Jet (Acosta & Huber, 2017),  
 147 bifurcating towards northwestern India and the Himalayan foothills.

148 The EASM is characterized by a band of strong convective precipitation stretching  
 149 from Myanmar to Japan along eastern China, also called the Meiyu-Bayu front. This high-  
 150 precipitation zone is fed by warm moist air, advected from the Bay of Bengal and nearby  
 151 Indo-Pacific Warm Pool (IPWP) (Fig. SI 14 d). The Meiyu-Bayu front results from the  
 152 ascension of these moist winds over the Jet Stream in the subtropics (Kong et al., 2017;  
 153 Molnar et al., 2010; Sampe & Xie, 2010). The progressive northward migration of the front  
 154 during spring and early summer is likely influenced by the seasonal displacement of the  
 155 Jet Stream relative to the Tibetan Plateau (TP): it is located south of the TP at  $\sim 25^\circ\text{N}$  in  
 156 winter, and progressively migrates to its northern edge in summer. The EASM, observed up  
 157 to  $\sim 40^\circ\text{N}$ , is an extra-tropical phenomenon, and is therefore regularly questioned as being  
 158 a "real" monsoon (Molnar et al., 2010; R. A. Spicer et al., 2016).

## 159 2.2 Winter season

160 Boreal winter is marked by a generally dry climate over most of the Asian continent.  
 161 North of  $\sim 20^\circ\text{N}$ , Eastern Asia is swept by northwestern winds bringing cold air and dusts  
 162 from inland Asia, while south of  $\sim 20^\circ\text{N}$ , northeastern winds carry moisture from the Pacific  
 163 Warm Pool, leading to monsoonal precipitation over Indonesia and northern Australia (Fig.  
 164 SI 14 f). The Indian subcontinent experiences a more complex wind pattern, as the westerly  
 165 winds in the North are channeled by the Zagros and Himalaya-Tibetan orography and then  
 166 deviated into northeasterly winds over the Arabian Sea, towards the eastern African coast  
 167 (Fig. SI 14 f). These peculiar winds lead to the formation of upwellings in the Bay of Bengal  
 168 and the Arabian Sea, allowing phytoplankton productivity blooms to occur in winter as well  
 169 (Lévy et al., 2007). This typical winter wind circulation (Fig. SI 14 f) is controlled by the  
 170 pressure gradients forming between the Siberian High and the surrounding Aleutian Low  
 171 (over the North Pacific) to the East and the low pressure center settled over the Maritime  
 172 Continent to the South (L. Wang & Chen, 2014). The spectacular extent and strength of  
 173 the Siberian High in boreal winter is a consequence of a strong radiative cooling of the lower  
 174 troposphere over snow-covered Asia (Jeong et al., 2011; Cohen et al., 2001; Jhun & Lee,  
 175 2004).

## 176 2.3 Summary: what traces leave the monsoons ?

177 Asian climate is under the crossed influence of the dry East Asian winter monsoon and  
 178 of three moisture sources, the westerlies and the East and South Asian summer monsoon  
 179 (Fig. 1). Nowadays, the boundary between regions receiving most of their annual precipita-  
 180 tion via the monsoons or via the westerlies is clearly bounded by the Himalayas in the South,  
 181 while it fluctuates in the East over China (Chen et al., 2021), crossing Eastern Tibet and  
 182 the Chinese Loess Plateau (Fig. 1). This boundary however likely evolved in the past, espe-  
 183 cially during the Paleogene when the orography and land-sea distribution were substantially  
 184 different from today (no Himalayas, less extent Tibetan Plateau, presence of the Paratethys  
 185 epicontinental Sea until the EOT) (X. Liu, Guo, et al., 2015) (see section 4). Tracking the  
 186 monsoons back in time can be done thanks to a variety of paleoclimatic archives recording  
 187 either their winds (intensity, direction) or precipitation (amount, seasonality) (P. Wang et  
 188 al., 2005; Zhisheng et al., 2015; Tada et al., 2016). Markers of aridification on the other  
 189 hand, such as increased proportion of xerophytic or halophytic plants or aeolian deposits  
 190 (dusts, loesses) in the fossil record are often interpreted as evidence for the presence of a  
 191 winter monsoon. The next section depicts the evolution of the Asian climate throughout  
 192 the Cenozoic based on these different paleoclimate indicators.

## 193 3 Increasingly old testimonies of paleo-monsoons through times

194 The classical view for the Asian paleoenvironment evolution throughout the Cenozoic  
 195 usually proposes a transition from a Paleogene "zonal climatic pattern" to a Neogene "mon-  
 196 soonal climatic pattern" (Guo et al., 2008; X. Sun & Wang, 2005; Jia et al., 2003). The

197 zonal pattern is described as a widespread arid to semi-arid band stretching across China  
198 and bracketed in its southern and northern borders by two humid belts, while the mon-  
199 soonal pattern hosts a core of pronounced aridity in inland Asia, surrounded by monsoonal  
200 (seasonally wet) climate on its southern and eastern edges. This view is mostly based on  
201 paleoclimate indicators compilations (pollen, fossils and lithological arguments, e.g. from  
202 Boucot et al. (2013)) used to infer either humid conditions (coal, indicators of forested en-  
203 vironments, large mammals, etc.) or arid ones (evaporites, desert and shrub environment,  
204 etc.). Recent evidence for highly seasonal climate in India, Myanmar and China as early  
205 as in the Eocene, at least 20 Myr earlier than previously thought, nevertheless challenge  
206 this paradigm. This section aims at presenting in the most comprehensive way the broad  
207 paleoenvironmental changes inferred from paleoclimate indicators in Asia from the Eocene  
208 to the early Pliocene, which are tentatively gathered in Fig. 2 (upper part). Localities and  
209 basins mentioned throughout the text are localized in Fig. 1.

### 210 **3.1 Eocene to Oligocene paleoenvironment evolution**

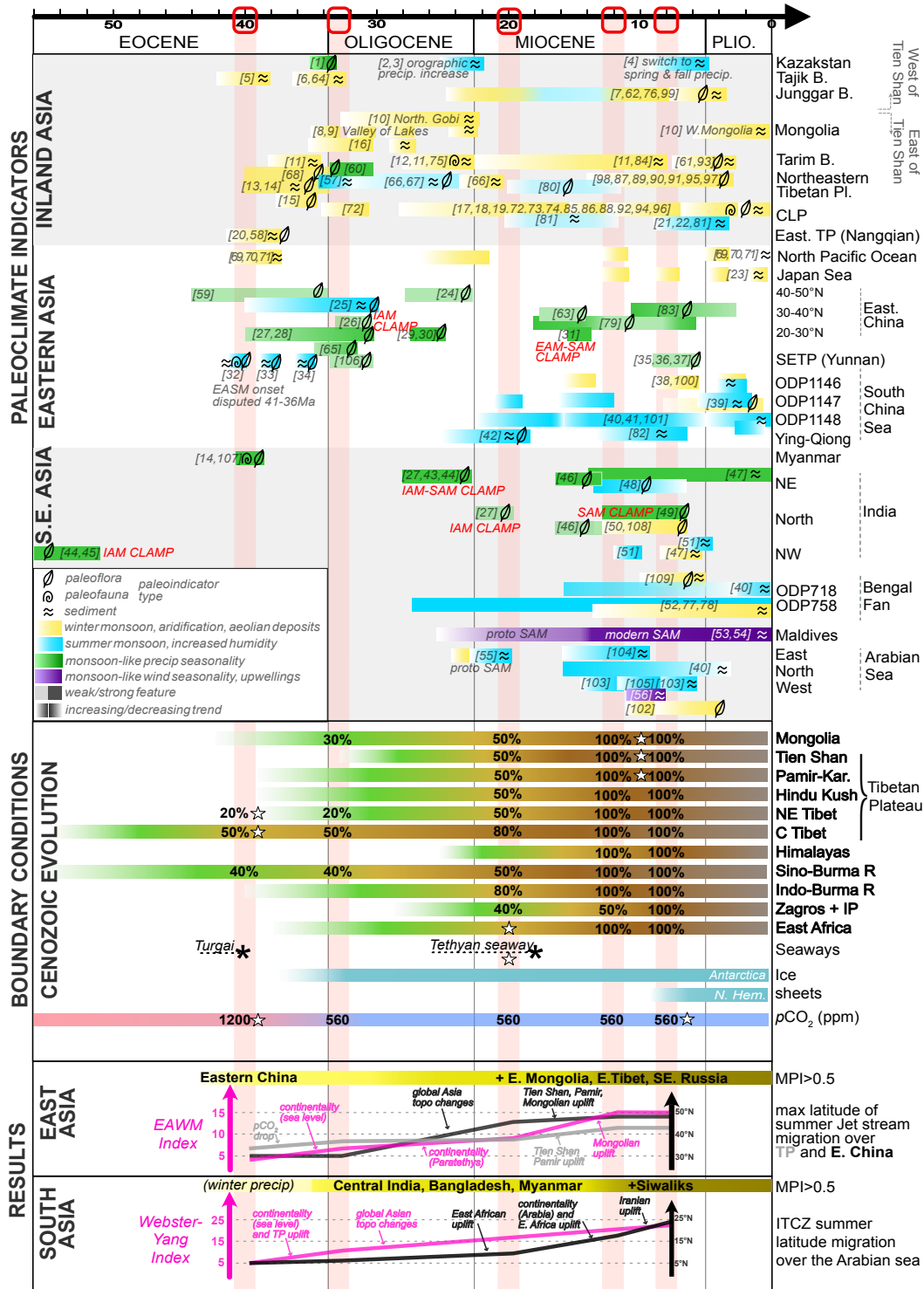
#### 211 *3.1.1 A progressive increase in summer monsoon indices from the Eocene* 212 *to the Oligocene*

213 Oldest testimonies of strongly seasonal precipitation in Southern Asia come from early  
214 Eocene floras situated in northwestern India (Shukla et al., 2014; Bhatia, Khan, et al., 2021)  
215 and from mid-late Eocene (40 Ma) floras and isotopic measurements in freshwater gastropod  
216 shells and mammals tooth enamel found in Myanmar (Licht et al., 2014; H. Huang et  
217 al., 2021). Both regions were however located further South at their respective period of  
218 deposition (between the equator and  $\sim 10^\circ\text{N}$ , Shukla et al. (2014); Westerweel et al. (2019))  
219 and their seasonality is therefore mainly interpreted within the scope of a pronounced ITCZ  
220 seasonal migration rather than monsoonal climate.

221 Further east, in the Yunnan region (Southwestern China), today submitted to a mix-  
222 ture of Southern and Eastern Asian summer monsoon, a shift from arid/semi-arid to humid  
223 environment is recorded during the late Eocene and interpreted as the onset of the Asian  
224 monsoon at that time (Sorrel et al., 2017; Fang et al., 2021; Zheng et al., 2022). Such cli-  
225 mate transition is inferred by the presence of coal layers, a change in the pollen record from  
226 xerophytic to mixed forest, change in lithology interpreted as lake and swamps expansion,  
227 together with the presence of fossils of freshwater fishes, gastropods and large mammals.  
228 While the precise date for this paleoenvironmental change is still debated between  $\sim 41$  Ma  
229 (Fang et al., 2021),  $\sim 36$  Ma (Zheng et al., 2022) and  $\sim 35.5$  Ma (Sorrel et al., 2017), it never-  
230 theless predates the Eocene-Oligocene Transition ( $\sim 34$  Ma). Proposed driving mechanisms  
231 to explain this regional climate shift usually involve either the crossing of a Tibetan Plateau  
232 elevation threshold (Zheng et al., 2022), a response to the Paratethys Sea retreat in the  
233 west, that would have allowed the penetration of more moisture from the east (Fang et al.,  
234 2021), or a response to the global climate cooling initiated after the Middle Eocene Climate  
235 Optimum (MECO) and subsequent sea surface temperature patterns reorganization (Sorrel  
236 et al., 2017).

237 Signs for the presence of an Eocene East Asian summer monsoon are also debated.  
238 In eastern China, Eocene paleovegetation mostly point to warm, humid and weakly sea-  
239 sonal evergreen forested environments on the coast (R. A. Spicer et al., 2016; X. Ma et al.,  
240 2012; Pound & Salzmann, 2017), except for northeastern middle-late Eocene Chinese floras  
241 retrieved from the coal deposits of the Huadian basin, that already seem to indicate adapta-  
242 tion to highly seasonal climate (Meng et al., 2018). Further inland, toward the eastern edge  
243 of the TP, these paleoenvironments gradually translate into sub-humid and/or seasonally  
244 dry conditions, as attested by various basin displaying alternating mudstone and evaporite  
245 deposits (D. Wang et al., 2013; Abels et al., 2011). Highly resolved deposits from the  
246 Jiangnan basin, spanning the mid Eocene to early Oligocene ( $\sim 40$ -34 Ma), together with  
247 modeling studies (Tardif et al., 2021), have shown that the penetration of moisture into east-





**Figure 2.** upper Paleoclimate tendencies recorded in inland, eastern and southern Asia. Bibliographic references in brackets are listed in SI Table 3 ; center Evolution of main landforms, seaways, ice sheets and pCO<sub>2</sub> during the Cenozoic overlain with values used in our reference simulations (elevations expressed in % of modern). White stars indicate parameters that were tested with sensitivity experiments ; bottom Overview of the main monsoon indicators evolution obtained in this study. We used the following abbreviations: NETP (SETP) - Northeastern (Southeastern) Tibetan Plateau, CLP - Chinese Loess Plateau.

ern China in the late Eocene was likely strongly modulated by orbital forcing. This may indicate that this Paleogene wet/dry zonal climatic pattern previously introduced (Guo et al., 2008; X. Sun & Wang, 2005) was already periodically disturbed in the late Eocene, and may have prefigured a more stable and widespread monsoon-like precipitation seasonality settlement in the Oligocene.

Indeed, after the global cooling and aridification occurring at the Eocene-Oligocene Transition, pieces of evidence for monsoon-like precipitation seasonality of variable intensity become more widespread in floras from both China (X. Ma et al., 2012; Herman et al., 2017; Vornlocher et al., 2021; J. Ren et al., 2021; Miao, 2013; Ling et al., 2021; S. Li et al., 2018; C. Huang & Hinnov, 2019; H. Tang et al., 2020) and India (Bhatia, Khan, et al., 2021; R. Spicer et al., 2017; Srivastava et al., 2012). Additionally, oceanographic records from adjacent seas (Bay of Bengal and South China Sea) that were recently extended back to the late Paleogene, have been used to track the SAM and EAM evolution from the middle Oligocene onward. Drilling from the Ying-Qiong Basin in the South China Sea, spanning the late Oligocene to early Miocene (25-18 Ma), record an increase in terrigenous organic matter influx, together with increased tropical-subtropical angiosperm pollen proportion, which are interpreted as signs of increasing East Asian summer monsoon during this period (W. Ding et al., 2021). In the Bay of Bengal (ODP 758), measurements of clays radiogenic isotopic composition, indicating undisturbed weathering patterns over the last 27 Ma, are interpreted as signs of stable South Asian summer monsoon activity since at least the late Oligocene (S. Ali et al., 2021).

### 3.1.2 Late Eocene aridification in response to the Paratethys Sea retreat and global climate cooling

Markers of arid to sub-humid climate are already widespread in inland Asia in the Eocene with, for example, many sites displaying pollen assemblages characteristics of steppe-desert environments such as in the Xining, Hoh Xil, Qaidam and Tarim basins (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 records an evolution toward drier conditions, as suggested by increased proportions of xerophytic plants such as *Nitraria* and *Ephedra* in Xining, Hoh Xil, Qaidam (Northeastern Tibet), Nangqian (central-eastern Tibet) and Tarim basins (Q. Yuan et al., 2020; Barbolini et al., 2020; Hoorn et al., 2012; Miao, Wu, et al., 2016),  $\delta^{18}\text{O}$  measurements from ostracod shells in the Tarim basin (Bougeois et al., 2018), clumped isotopes from pedogenic carbonates in the Xining basin (Page et al., 2019), compound-specific hydrogen isotope analyses ( $\delta^2\text{H}$ ) applied to sedimentary leaf wax *n-alkanes* in the Qaidam basin (Wu et al., 2021) and by lipid biomarkers and compound-specific carbon isotopic compositions from bulk sediments in the Nangqian basin (J. Wei et al., 2022). The presence of detrital material attributed to Asian dusts is also recorded in sediments from Central North Pacific (GPC3 and ODP site 1215) since ~40 Ma (Pettke et al., 2002; D. Rea et al., 1985; Ziegler et al., 2007).

This late Eocene aridification stage is mostly interpreted as resulting from the reduced westerly moisture input during the two successive Paratethys Sea regression phases from the Tarim and the Tajik basins, recorded 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). Highly resolved sediment deposits, such as those of the Xining basin (northeastern Tibet) display good correlation between wetter evaporite layers and Paratethys Sea transgressions phases into the Tajik and Tarim basins, and between arid mudstone intervals and regression phases. This confirms the key role played by the westerlies and the Paratethys Sea as a major moisture source for inland Asia in the Eocene, even over long distances (Meijer et al., 2019; Bosboom et al., 2014a). Additionally, a strong obliquity cyclicity is imprinted on the Xining evaporite/mudflat layers in the late Eocene (40-34 Ma), coeval with the appearance of loess-like dusts in the mudflat phases (Meijer et al., 2021; Licht et al., 2014). This hints a marked influence of high latitudes dynamics, such as incipient ice-sheets at the poles (Abels et al.,

2011; Xiao et al., 2010) and variations in the Siberian High intensity (Meijer et al., 2021), on inland Asian climate at that time.

### 3.1.3 *Oligocene climate fluctuations and inception of major modern Asian deserts*

The Oligocene evolution of inland Asian climate appears both regionally and timely contrasted. In the early Oligocene, a few sites situated within the westerly moisture region tend to show increased humidity trends, recorded by  $\delta^{18}\text{O}$  isotopic measurements in pedogenic carbonates (Kent-Corson et al., 2009) in the Tarim and western Qaidam basins together with  $\delta^2\text{H}$  measurements in leaf waxes (Wu et al., 2021) and leaf fossil CLAMP analysis suggesting wet environment with weak (not monsoonal) precipitation seasonality (Song et al., 2020) from western Qaidam. On the other hand, many testimonies for drier climate are reported within the eastern Asian monsoon domain by  $\delta^{18}\text{O}$  isotopic measurement in pedogenic carbonates from the eastern Qaidam basin (Y. Sun et al., 2020; Kent-Corson et al., 2009), Lanzhou basin in the Chinese Loess plateau area (B. Li et al., 2016), together with CLAMP analysis on fossil leaves advocating for an increased (and monsoon-like) precipitation seasonality in Lanzhou (Miao, 2013), and Markam basin in southeastern TP (T. Su, Spicer, et al., 2019), in agreement with records of increased monsoon-like precipitation seasonality in Eastern China introduced in the previous section. Interpretations nevertheless diverge and suggest that this aridification trend may indicate either weakened East Asian summer monsoon precipitation due to global cooling after the EOT (Wu et al., 2021), increased winter aridity due to the final Paratethys Sea retreat, global sea level fall after the EOT and/or global cooling (Miao, 2013; B. Li et al., 2016), or be due to TP uplift (T. Su, Spicer, et al., 2019).

In the Taatsin Gol region, at the southern edge of the Hangay Dome and extreme north of the modern Gobi desert, isotope measurement from pedogenic carbonates indicate aridification since the early Oligocene (Caves Rugenstein et al., 2014). Finer resolved records reveal several bursts of aridification at  $\sim 34$ -33,  $\sim 31$ ,  $\sim 28$  and  $\sim 23$  (Baldermann et al., 2021) and the presence of loess as soon as  $\sim 34$  Ma (J. Sun & Windley, 2015). Proposed mechanisms for such aridification involve increased continentality due to land-sea distribution fluctuations at 34 Ma (J. Sun & Windley, 2015) and an early Hangay Dome uplift that would have acted as a barrier to moisture advection from Siberia to Mongolia as soon as the Oligocene (Caves Rugenstein et al., 2014). Better resolved records however highlight the good correlation in time between ice sheets expansion phases and these aridification pulses, and therefore propose that large scale climate fluctuations at that time were the main drivers of Mongolia hydrologic dynamics (Baldermann et al., 2021). Further south, in the Bayanhot basin (southern Gobi), two bursts of aridification inferred from environmental magnetic, mineralogical and geochemical study at  $\sim 31$  and  $\sim 28$  Ma seem to further confirm the influence of high latitude ice sheet fluctuation on Mongolian Oligocene climate (Wasiljeff et al., 2022).

In the mid to late Oligocene, massive eolian dust deposits are reported in three major sites: in the Chinese loess plateau in Northeastern Tibet starting at  $\sim 29$  Ma (Garzzone et al., 2005; Qiang et al., 2011) (see (Meijer et al., 2021) for a full review), in the Tarim Basin (hosting today the Taklamakan desert) at  $\sim 27$ -22 Ma (Zheng et al., 2015) and in the Junggar basin at  $\sim 24$  Ma (J. Sun et al., 2010). Aridification is also inferred from  $\delta^{18}\text{O}$  isotopic measurements from ostracods and pedogenic carbonates from the Tarim basin at  $\sim 25$  Ma (Bougeois et al., 2018; Kent-Corson et al., 2009) and from isotopic records in stromatolites from the Junggar basin loosely dated in the early Miocene (W. Yang et al., 2019). This second step of aridification is coeval with  $\delta^{18}\text{O}$  and  $\delta^{13}\text{C}$  isotopic measurements from pedogenic carbonates suggesting increased orographic precipitation in the Issyk Kul and Ili basins, both situated on the western side of the Tian Shan ranges (Hellwig et al., 2018; Macaulay et al., 2016). Collectively, these observations are interpreted as the sign that the Tian Shan-Pamir uplift had reached an elevation threshold that would have started



shielding the Tarim and Junggar basins from westerlies moisture input, thus favoring their aridification (Bougeois et al., 2018; W. Yang et al., 2019).

This late Oligocene to early Miocene aridification phase seems to be recorded by several oceanic drillings, for example in Central North Pacific, with an increase in dust mass accumulation rate recorded at  $\sim 25$  Ma by (drilling site GPC3) (D. K. Rea, 1994; Pettke et al., 2002) and at  $\sim 22$  Ma (ODP site 1215) (Ziegler et al., 2007). These drillings are however situated far from the Asian coast and underwent important plate motion throughout the Cenozoic, making it difficult to disentangle variations due to inland Asia climate changes, possible global atmospheric/oceanic circulation patterns changes, and variations in the recorded signal due to plate motion. The presence of a proto winter monsoon wind circulation is also suggested by increased proportion of dusts, inferred from magnetic susceptibility measurement in the Maldives (IODP sites U1467, U1468) at  $\sim 24$  Ma (Betzler et al., 2016), and by an increase in water column mixing and ventilation evidenced by a reduction in  $\Delta \delta^{13}\text{C}$ , increase in bulk sediment Mn/Fe and Mn/Ti and foraminiferal Mn/Ca in the eastern Arabian Sea (Beasley et al., 2021).

## 3.2 Strengthening of summer and winter monsoon patterns in the Neogene

### 3.2.1 *An East-West gradation in Miocene South Asian summer monsoon variations and the expansion of grassland*

Despite a better coverage in both continental and marine records in the Neogene, substantially diverging interpretations remain regarding the evolution of the monsoons winds and precipitation seasonality, especially during the late Miocene. Precipitation proxies from the South Asian Monsoon domain allow to highlight an East-West gradation in climate evolution throughout the Miocene, recorded in the Siwaliks deposits of the Himalayan foreland basin. In northeasternmost India (Arunachal Pradesh), modern-like monsoon precipitation seasonality is inferred from plant macrofossils since the middle Miocene (13 Ma) and displays little variation in intensity when compared to early Pleistocene fossils from the same locality (Khan et al., 2014). Further west, plant macrofossils and pollen from Darjeeling, Nepal, northwestern India and Pakistan show a progressive replacement of middle Miocene evergreen forests by deciduous forests in the late Miocene, and finally a massive expansion of grassland in the latest Miocene (Hoorn et al., 2000; Bhatia et al., 2022; Bhatia, Srivastava, et al., 2021; Srivastava et al., 2018). This late Miocene floristic change is also accompanied by several shifts in  $\delta^{13}\text{C}$  toward more positive values, recorded between 10.5 and 6 Ma by isotopic measurements in pedogenic carbonates and organic matter and lipid biomarkers in the Central Siwaliks (Nepal) (Quade et al., 1995; Neupane et al., 2020; Dettman et al., 2001), and in the Western Siwaliks (northwestern India and Pakistan) (Sanyal et al., 2004; Vögeli et al., 2017; Quade & Cerling, 1995). Similar trends are also recorded in nearby marine basins by isotopic measurements from bulk sediments and leaf wax, pollen and macrofossils at  $\sim 8$ -7 Ma in the Arabian Sea (ODP 722 and IODP U1457) (Feakins et al., 2020; Y. Huang et al., 2007) and at  $\sim 7$ -6 Ma in the Bengal fan (Polissar et al., 2021).

This marked carbon isotope excursion is observed in other localities of the globe and testify of an important ecological change, characterized by the replacement of plants using the C3 photosynthetic pathway by grasslands using the C4 photosynthetic pathway (Cerling et al., 1993; Tauxe & Feakins, 2020; Edwards et al., 2010). Recent re-calibration of these Asian continental carbon isotope records and comparison to other available records allow to show that C4 expansion occurred at distinct times:  $\sim 10$  Ma in Africa,  $\sim 8$  Ma in Pakistan,  $\sim 7$  Ma in peninsular India,  $\sim 6$  Ma in Nepal and later during the Pliocene in Australia and South America (Tauxe & Feakins, 2020). Although a late Miocene  $p\text{CO}_2$  drawdown was initially proposed as the main driver for the C3/C4 transition, such distinct timing in ecological turnover over the globe is assumed to reflect the multiplicity of forcing factors at play in this major biome reorganization. In the Siwaliks, this East-West climate gradation likely reflects the crossing of a precipitation threshold somewhere in Nepal (Central Siwa-

403 liks), with northwestern India, Nepal and Pakistan experiencing drier winters and overall  
404 aridification in the late Miocene compared to the more stably humid northeastern Indian re-  
405 gion. The C4 grasslands expansion would thus possibly result from a combination of  $p\text{CO}_2$   
406 decrease, reduced moisture advection from the westerlies in this region, Tibetan Plateau  
407 and Himalayas uplift and to more frequent forest fire that would have favored grasses over  
408 trees (Sanyal et al., 2004; Vögeli et al., 2017; Quade & Cerling, 1995; Srivastava et al., 2018;  
409 Tauxe & Feakins, 2020; Polissar et al., 2021; Feakins et al., 2020).

410 Precipitation proxies recovered from marine drillings offer contrasting information. In  
411 the central Bay of Bengal (ODP site 758), relatively undisturbed isotopic composition of  
412 clays issued from silicate weathering since the Oligocene ( $\sim 27$  Ma) are interpreted as signs  
413 of a stable South Asian summer monsoon presence since that time (S. Ali et al., 2021).  
414 On the other hand, chemical weathering (CIA, K/Al ratio) and sediment accumulation  
415 rates from both the Indus Fan and the Bay of Bengal recording the physical erosion of the  
416 Himalayas and southern Tibet, suggest that SASM precipitation reached a maxima in the  
417 middle Miocene ( $\sim 10$ -16 Ma) before declining between  $\sim 8$ -3 Ma (P. Clift et al., 2008). This  
418 decline was primarily interpreted as reflecting a weakened SASM precipitation (P. Clift et  
419 al., 2008), although other studies based on similar proxies from the Bay of Bengal argue it  
420 may rather reflect an intensification of the winter monsoon (S. Ali et al., 2021), possibly  
421 coupled to a weakening of the South Asian summer monsoon (J. Lee et al., 2020), which  
422 seem in better agreement with continental proxies, and with marine records from the South  
423 China Sea (Wan et al., 2010) (developed in the next section).

424 Wind proxies retrieved from the Arabian Sea and the Bengal Fan pointing to the  
425 presence of seasonal wind reversal (hence of a winter monsoon) since the middle Miocene  
426 seem to corroborate this explanation, although some aspects of the SAM wind circulation  
427 evolution remain debated. In the Arabian Sea, the appearance of *Globigerina Bulloides*  
428 in the sediments had initially been used to infer an intensification (or an onset) of the  
429 South Asian summer monsoon wind circulation at  $\sim 8$  Ma (Kroon et al., 1991), based on  
430 the assumption that stronger SASM winds would translate into stronger upwelling activity.  
431 Those records have since then been extended back in time (Bialik et al., 2020; Zhuang et al.,  
432 2017; Gupta et al., 2015) and combined with analysis of current-controlled drift sediments  
433 and geochemical tracers from the Maldives archipelago (Betzler et al., 2016). These new  
434 studies now point to a settlement of a "proto" South Asian monsoon wind circulation since  
435 the late Oligocene ( $\sim 25$  Ma), and of its reinforcement in the middle Miocene ( $\sim 13$ -10 Ma)  
436 (Betzler et al., 2016; Zhuang et al., 2017; Bialik et al., 2020) and possibly also in the late  
437 Miocene ( $\sim 8$ -7 Ma) (Gupta et al., 2015). Nevertheless, increasing number of studies based  
438 on marine records from both the Arabian Sea and the Bengal Fan do not observe this latest  
439 Miocene (8-7 Ma) SASM peak (Bolton et al., 2021; Tripathi et al., 2017; Betzler et al., 2016;  
440 Y. Huang et al., 2007). Strong SAM wind circulation since at least 10 Ma is also supported  
441 by clay mineralogy, detrital isotopes, export productivity and sediments accumulation from  
442 records in the southern Bay of Bengal (IODP 1443) (Bolton et al., 2021; J. Lee et al., 2020))  
443 and the eastern Arabian Sea (IODP 1456) (Tripathi et al., 2017)).

### 444 3.2.2 *EASM in the Neogene*

445 Summer precipitation typical of the East Asian summer monsoon is assumed to occur  
446 throughout the Neogene (22-3 Ma), as attested by sediments exported to the South China  
447 Sea, such as weathering indices (P. Clift et al., 2008),  $\delta^{13}\text{C}$  measurements in black carbon  
448 (Jia et al., 2003), pollen, biomarkers, kerogen composition and clay mineralogy (W. Ding et  
449 al., 2021). On land,  $\delta^{18}\text{O}$  from paleosol carbonates in the CLP (Suarez et al., 2011) and most  
450 Chinese floras, studied with Coexistence Approach, CLAMP or Bayesian method applied  
451 to pollen, point to summer precipitation comparable to modern amounts (Q. Wang et al.,  
452 2021; Yao et al., 2011; Xing et al., 2012; Hui et al., 2021; B. Wang et al., 2021; Q.-g. Sun et  
453 al., 2002). Most of those studies nevertheless suggest wetter winters due to weaker winter  
454 monsoon until the latest Miocene, with a few exceptions, such as the mid-Miocene floras

455 from the Zhangpu biota in southeastern China characteristic of megathermal seasonal forest  
 456 which CLAMP signature showing affinities with EAM and SAM floras (B. Wang et al., 2021).  
 457 The variations in EASM intensity throughout the Miocene remain controversial in both  
 458 continental and marine deposits, but two intensification peaks usually emerge, tentatively  
 459 linked to the Middle Miocene Climatic Optimum (~17-14 Ma) and to the Miocene-Pliocene  
 460 boundary (~8-5 Ma).

461 The middle Miocene EASM peak is inferred from sediments from the South China Sea  
 462 by chemical weathering proxies at 15-12 Ma (P. Clift et al., 2008), and  $\delta^{13}\text{C}$  in black carbon  
 463 (Jia et al., 2003), although the latter rather point to a step-wise intensification of EASM in  
 464 the course of the Neogene (at ca. 20, 15, 10, 6 and 2 Ma). On land, characterization of iron  
 465 oxides from the Chinese Loess Plateau suggest an EASM peak at 16-14 Ma (H. Zhao et al.,  
 466 2020). The proportion of *Fupingopollenites*, an extinct palynomorph supposed to be highly  
 467 dependent on summer moisture, has also been used in an extensive compilation spanning  
 468 available Miocene Chinese pollen records and point to an important penetration of EASM  
 469 moisture, up to the eastern Qaidam basin, during the MMCO, before retreating to the East  
 470 in the rest of the Neogene (Miao, Song, et al., 2016).

471 A substantial increase in EASM strength is also robustly suggested by many lines of  
 472 evidence in the late Miocene to Pliocene, although with notable variations in onset timing  
 473 (10-4 Ma). In the South China Sea, it is inferred from  $\delta^{18}\text{O}$  and Mg/Ca derived temper-  
 474 ature records of surface and sub-surface water (IODP U1501) (C. Yang et al., 2021), clay  
 475 minerals, grain size, major and trace elements analysis from terrigenous sediments (ODP  
 476 1146) (Wan et al., 2007, 2010) and by black carbon  $\delta^{13}\text{C}$  (Jia et al., 2003). In China, this  
 477 EASM intensification is recorded through mineralogical, chemical and magnetic analysis of  
 478 sedimentary deposits in the Chinese Loess Plateau (Ao et al., 2016, 2021; H. Zhao et al.,  
 479 2020), and by pollen from the Tianshui (Hui et al., 2021) and the Weihe Basin (L. Zhao  
 480 et al., 2020). The forcing factors responsible for these EASM intensification peaks remain  
 481 debated, but are usually attributed to global warming (Ao et al., 2021; H. Wang et al.,  
 482 2019) and/or to Neogene TP uplift phases (Hui et al., 2021; X. Ren et al., 2020; H. Zhao  
 483 et al., 2020). Although EASM and EAWM evolution show few correlation during most of  
 484 Cenozoic, this late Miocene EASM increase is notably accompanied with EAWM increase  
 485 (described in next section).

### 486 **3.2.3 Miocene to Pliocene aridification pulses in response to regional up-** 487 **lifts**

488 Throughout the Miocene, dust deposition continues in the Chinese Loess Plateau (Guo  
 489 et al., 2002; Meijer et al., 2021), Tarim-Taklamakan (Kent-Corson et al., 2009; Zheng et  
 490 al., 2015; Heermance et al., 2018) and Junggar basins (J. Sun et al., 2010). In parallel  
 491 with previously mentioned indices for an increased EASM during the MMCO (~17-14 Ma)  
 492 (P. Clift et al., 2008; Jia et al., 2003; H. Zhao et al., 2020), a warmer and less arid climate  
 493 interval is suggested by pollen showing relatively lower proportion of xerophytic taxa in the  
 494 Junggar (J. Sun et al., 2010) and Qaidam basins (Miao et al., 2011). After the MMCO,  
 495 a pronounced step-wise aridification is recorded in most of inland Asia until the Miocene-  
 496 Pliocene boundary (Lu et al., 2010; Lu & Guo, 2014; Z.-H. Tang & Ding, 2013). Here again,  
 497 the timing for aridification steps varies with the region and is usually tentatively explained  
 498 by a combination of global cooling, ice-sheet expansion and multiple regional uplift episodes  
 499 occurring at that time.

500 In the eastern part of inland Asia, encompassing the Chinese Loess Plateau, central  
 501 China and northeastern Tibetan Plateau, this aridification is attributed to northeastern TP  
 502 uplift (Miao et al., 2012), Altai mountains uplift (Caves Rugenstein et al., 2014) and/or late  
 503 Miocene cooling (Lu et al., 2010; Lu & Guo, 2014; Peng et al., 2016). Such aridification is  
 504 supported by overwhelming evidences, such as pollen and biomarkers documenting a step-  
 505 wise transition from sub-humid to arid environment starting at ~15 Ma (J. Liu et al., 2016;

506 Miao et al., 2011; Jiang & Ding, 2008; Peng et al., 2016), massive grasses expansion at  
 507  $\sim$ 11-7 Ma (Suarez et al., 2011; Barbolini et al., 2020; Y. Ma et al., 2005; L. Zhao et al.,  
 508 2020; H. Wang et al., 2019), widespread eolian deposition at  $\sim$ 8-7 Ma (X. Ma & Jiang, 2015;  
 509 Guo et al., 2002; Qiang et al., 2011; B. Li et al., 2016; Jiang et al., 2017), increased  $\delta^{18}\text{O}$   
 510 of pedogenic carbonates and terrestrial mammals tooth enamel at  $\sim$ 14-13 and  $\sim$ 7-5 Ma  
 511 (B. Li et al., 2016; Kent-Corson et al., 2009; Wu et al., 2021; W. Liu et al., 2014; Y. Sun et  
 512 al., 2020; Kaakinen et al., 2006; Y. Wang & Deng, 2005), *n*-alkanes analysis from Qaidam  
 513 sediments since 13 Ma (Z. Liu et al., 2014) and dominance of cold-aridiphilous mollusks  
 514 proportion in the CLP from 7-5 Ma (F. Li et al., 2008).

515 The western part of inland Asia, comprising the Tarim and Junggar basins and several  
 516 sites in eastern Kazakhstan, which all add to the body of evidence indicating mid to late  
 517 Miocene aridification, as shown by increased xerophytic and grasses pollen proportions at  
 518  $\sim$ 7-5 Ma (J. Sun et al., 2010; J. Sun & Zhang, 2008; Z. Zhang & Sun, 2011; Barbolini et  
 519 al., 2020; J. Sun et al., 2008), massive eolian deposition in the western Tarim at  $\sim$ 12-7 Ma  
 520 (Heermance et al., 2018) and permanent drying of lakes in the eastern Tarim at  $\sim$ 5 Ma  
 521 (W. Liu et al., 2014). The pedogenic carbonates  $\delta^{18}\text{O}$  record of these regions nevertheless  
 522 offers diverging pattern evolution, that have been tentatively explained by their location  
 523 with respect to the Tian Shan and Altai mountains (Caves Rugenstein et al., 2017). Regions  
 524 today lying upwind of the Tian Shan and Altai mountains, such as the Issyk Kul (Macaulay  
 525 et al., 2016), Junggar (Charreau et al., 2012) and Zaysan basins (Caves Rugenstein et  
 526 al., 2017) show a decline in  $\delta^{18}\text{O}$  between 10-5 Ma (varying from  $\sim$ 1.5 to 4 ‰) which  
 527 was interpreted as a change in precipitation seasonality following the uplift of the Altai  
 528 Mountains (Caves Rugenstein et al., 2017). On the other hand, increased  $\delta^{18}\text{O}$  values since  
 529  $\sim$ 15 Ma and peaking at  $\sim$ 7-5 Ma in the Tarim basin, situated downwind of the Tian Shan  
 530 ranges display trends similar as those of the Qaidam basin (Kent-Corson et al., 2009), which  
 531 are interpreted as indicating aridification at that time.

532 EAWM intensification peaks are also recorded offshore at  $\sim$ 15-12,  $\sim$ 8-5 and  $\sim$ 3 Ma,  
 533 in usually good coherence with continental data. In the South China Sea, it is suggested  
 534 by increased black carbon concentration and accumulation (whose transportation to the  
 535 Sea would be favored by stronger winter winds) (Jia et al., 2003), planktic and benthic  
 536  $\delta^{18}\text{O}$  records and mixed layer temperature (A. Holbourn et al., 2021; A. E. Holbourn et al.,  
 537 2018), clay mineralogy and grain size analysis (Wan et al., 2007), and by indices of decreased  
 538 chemical weathering (G. Wei et al., 2006). In the Sea of Japan, clay mineral assemblage  
 539 and isotopic analysis of silicate fraction covering the last 15 Ma suggests a stepwise drying  
 540 of Central Asia suggested at 12, 8 and 3.5 Ma (Shen et al., 2017). In the North Pacific,  
 541 an increase in Asian dusts proportion is recorded at  $\sim$ 10 and  $\sim$ 4 Ma (D. Rea et al., 1985;  
 542 Pettke et al., 2002). Last, brutal onset of SAM wind seasonality (and therefore occurrence  
 543 of winter monsoon reversed winds) is inferred at  $\sim$ 12 Ma in the Maldives (Betzler et al.,  
 544 2016) and in the eastern Arabian Sea (X. Yang et al., 2020).

### 545 3.3 Assessing forcings and mechanisms through climate models

#### 546 3.3.1 SAM forcings assessed by modeling experiments derived from mod- 547 ern geography

548 Major forcings proposed to explain multi-million year monsoon variability are global  
 549 climate variations, changes in topography and changes in land-sea distribution. Attributing  
 550 a specific change to one or several of these forcings is nevertheless complicated by the  
 551 fact that they may vary at the same time, produce non-linear responses with potentially  
 552 antagonizing effects and because they are sometimes poorly dated or quantified. Modeling  
 553 experiments, first derived from modern geographies and boundary conditions, have helped  
 554 better understanding the role played by these different forcings in shaping the regional  
 555 monsoons.

556 The Tibetan Plateau has long been seen as instrumental in driving the South Asian  
557 monsoon circulation and precipitation, its elevated heated surface acting as a "heat pump"  
558 driving the convergence of the surrounding air masses (Kutzbach et al., 1993; Zou et al.,  
559 2019; Molnar et al., 1993). It seems now that it is rather the physical barrier created by  
560 the landform (the Himalayas or the Tibetan Plateau) that is essential to induce abundant  
561 orographic precipitation over northern India, through steering and lifting of moist air masses  
562 (Abbott et al., 2016; Boos & Kuang, 2010; Acosta & Huber, 2020). The condensation  
563 of moisture in these ascending air masses leads to important latent heat release in the  
564 high troposphere over northern India, which acts as a positive feedback sustaining summer  
565 convection and precipitation in South Asia (He, 2017). The large-scale monsoon circulation  
566 seasonal reversal resulting in the initial advection of these moist air masses over Asia in late  
567 spring and early summer, on the other hand, is essentially driven by surface ocean-continent  
568 temperature gradients and resulting pressure patterns (Acosta & Huber, 2020; Merlis et al.,  
569 2013).

570 Smaller orographic features were also shown to play a critical role in shaping and  
571 strengthen the South Asian monsoon. By channeling the Somali Jet and insulating it from  
572 the subtropical dry westerly flow, the Anatolian-Iranian landform were shown to greatly con-  
573 tribute to enhance moisture transport towards India and Eastern Asia (Acosta & Huber,  
574 2020; He, 2017; H. Tang et al., 2013). The Eastern African highlands, positioned on the way  
575 of the ITCZ flow, are critical in shaping both African and South Asian climate (Bannon,  
576 1979; Rodwell & Hoskins, 1995; Sepulchre et al., 2006). Recent studies have specifically  
577 demonstrated that Eastern African highlands presence contribute greatly to strengthen and  
578 concentrate the Somali Jet, but that their absence would result, quite counter-intuitively,  
579 into higher summer precipitation over SAM region, due to increased advection directly from  
580 the ocean to the continent (H.-H. Wei & Bordoni, 2016; Chakraborty et al., 2009). This last  
581 aspect hints that strong upwellings, due to strong atmospheric circulation, may not correlate  
582 with strong precipitation on land. The effect of the Pamirs, Tian Shan and northeastern  
583 TP uplifts on SAM remains unclear. When these uplift together with Mongolian landforms,  
584 H. Tang et al. (2013) simulate increased moisture advection to eastern China but decreased  
585 precipitation over India, while the opposite effect is obtained when not modifying (mod-  
586 ern) Mongolian topography (R. Zhang et al., 2017). Apart from differences coming from  
587 the model and boundary conditions, this likely points to the complex interactions existing  
588 between the SAM and EAM regional monsoons.

### 589 *3.3.2 EAM forcings assessed by modeling experiments derived from mod-* 590 *ern geography*

591 In contrast with the modest effect it has on the SAM, the Tibetan Plateau uplift is  
592 instrumental for the EASM, as the heating of the Plateau, induces a cyclonic circulation  
593 anomaly promoting advection of moist air masses towards Eastern Asia (Z. Zhang et al.,  
594 2007; H. Tang et al., 2013; R. Zhang et al., 2017). Additionally, simulations testing the  
595 effects of increased Asian continentality (to mirror a global sea level fall or the retreat of  
596 the Paratethys Sea, for example), have shown that it results in amplified summer moisture  
597 convergence from the Indian Ocean towards Asia, due to changes in surface temperature  
598 gradients deepening the continental low pressure belt (Z. Zhang et al., 2007).

599 Although smaller than the HTP complex, the Pamir, Tian Shan and Mongolian land-  
600 forms are adequately situated in latitude to interfere both dynamically and thermally with  
601 planetary-scale atmospheric circulation. Their uplift was shown to enhance the aridity in  
602 inland Asia (actual Gobi and Taklimakan deserts), as it reinforces the EAWM circulation  
603 patterns (X. Liu & Yin, 2002; X. Liu, Sun, et al., 2015; Baldwin & Vecchi, 2016; Sato,  
604 2009). By creating a cold pool and deviating the westerlies towards northern Siberia, the  
605 Mongolian orography has been proposed to be responsible for the Jet Stream northward  
606 migration in summer and intensity in winter (White et al., 2017; Shi et al., 2015; Sha et al.,  
607 2020), and for most of the Siberian High location and intensity (Sha et al., 2015).



### 608 **3.3.3 Modeling experiments derived from paleo boundary conditions**

609 Modeling studies based on modern geographies are nevertheless limited in to what  
 610 extent they allow to understand past changes, especially for periods where very different  
 611 geographies and climate prevailed. Paleoclimate simulations, using sets of realistic (as much  
 612 as possible) paleo-boundary conditions are however rarer. Indeed, they usually imply high  
 613 computational cost, and require important background knowledge on past constraints (pa-  
 614 leogeography, CO<sub>2</sub>, solar constant, ice-sheets volume).

615 More specifically, paleoclimate modeling has confirmed that increasing Asian conti-  
 616 nentality during the Oligocene and the Miocene, mirroring either the Paratethys Sea retreat  
 617 and/or global eustatic sea level fall, was a key leverage of increased moisture advection  
 618 toward Asia in summer (Fluteau et al., 1999; Ramstein et al., 1997; Z. Zhang et al., 2014;  
 619 Sarr et al., 2022). In Oligocene conditions, the uplift of the peripheral Tian Shan, Pamir  
 620 and northeastern portions of the TP was also shown to promote inland Asian aridity, when  
 621 compared to the sole uplift of the "core" Tibetan plateau (R. Zhang et al., 2017). In late  
 622 Eocene conditions, the latitude of the Tibetan Plateau was also shown to impact signif-  
 623 icantly and non-linearly East Asian climate: with a TP uplifting in the tropics ( $\sim 11^\circ\text{N}$ )  
 624 inducing precipitation on the reliefs but aridification in most of China, while a TP uplifting  
 625 at its modern latitude ( $\sim 26^\circ\text{N}$ ) would increase moisture advection to eastern China and  
 626 aridification of inland Asia (R. Zhang et al., 2018).

627 Although paleoclimate studies offer contrasted views regarding Eocene climate, some  
 628 suggesting the presence of monsoons at that time (Huber, 2003) and others not (Tardif et  
 629 al., 2020; Z. Zhang et al., 2012; R. Zhang et al., 2018), this divergence is most likely due to  
 630 the very different paleogeographic reconstructions that were used. Indeed, as many features  
 631 of the Eocene paleogeography remain highly controversial (height of the incipient Tibetan  
 632 topography, shape of Indo-Asian collision zone, height of peripheral landforms, land-sea  
 633 distribution, etc.), the geographic reconstructions may diverge widely from one study to  
 634 another. Paleoclimate modeling have nevertheless allowed to emphasize the prevalence of  
 635 geography impact on climate, rather than  $p\text{CO}_2$  variations or far-field ice-sheets effects  
 636 when it comes to the SAM and EAM settlement and intensification over long time scales  
 637 (Farnsworth et al., 2019; Thomson et al., 2021). Collectively, these information rather  
 638 reinforce the need for good paleogeographic constraints, and for more sensitivity experiments  
 639 in paleo context, which we aim at doing in the present contribution.

### 640 **3.4 Summary: a diachronous settlement of SAM and EAM seasonal fea-** 641 **tures, potential forcings and mechanisms and remaining uncertainties**

642 The past decade have brought plenty of new analysis based on fossil material, both  
 643 onshore and offshore. The history of SAM, EASM and EAWM settlement and intensification  
 644 now clearly seem to have been diachronous and its evolution tightly linked to land-sea mask,  
 645 topography and global climate evolution over long time scales. Some questions nevertheless  
 646 remain to be answered.

647 First, it is unclear whether Paleogene records of highly seasonal precipitations at low  
 648 latitudes ( $<20^\circ\text{N}$ ) do describe an actual monsoon, given that no wind proxies can be pro-  
 649 vided at this period, and that Eocene paleoclimate studies propose widely diverging scenar-  
 650 ios. CLAMP (Climate Leaf Analysis Multivariate Program) paleobotanical analysis (Feng  
 651 & Poulsen, 2016; Wolfe, 1993; Jacques, Su, et al., 2011) have brought interesting insight  
 652 on this point. Generally, they show that, although strong precipitation seasonality was  
 653 likely present since the Paleogene in broad parts of Southeastern Asia (Bhatia, Khan, et  
 654 al., 2021; R. Spicer et al., 2017; Herman et al., 2017), these floras did not yet developed  
 655 the characteristics of plants growing today in modern SAM or EAM regions until the mid-  
 656 dle Miocene ( $\sim 13$  Ma) (R. Spicer et al., 2017; Bhatia, Srivastava, et al., 2021). Instead,  
 657 Paleogene Asian floras display a signature typical of the regions today dominated by the  
 658 ITCZ seasonal migrations (referred to as Indonesian-Australian monsoon, or IAM in Fig-

659 ure 2). The reorganization from an ITCZ dominated to a monsoonal wind circulation may  
 660 have initiated in the Oligocene ( $\sim 28\text{-}23$  Ma), according to fossils from northeastern India  
 661 displaying a mixture of SAM and IAM leaf signatures (R. Spicer et al., 2017). These typical  
 662 monsoonal signatures likely emerged from a combination of both biotic and abiotic factors,  
 663 although CLAMP method is unable to disentangle the forcings responsible for these spe-  
 664 cific signatures. They however add to the body of evidence suggesting that precipitation  
 665 seasonality and typical monsoonal wind circulation may have been decoupled in the past.

666 Second, although modeling studies have started to explore localized landforms effect  
 667 in paleo context, there is plenty of room for improvement. For example, many of these  
 668 landforms have been tested collectively (eg. Tian Shan and Mongolia uplifting together),  
 669 and analyzed within the scope of a single time period. Significant hypotheses remain to  
 670 be tested in a paleo context, such as identifying the leverage for a Siberian High pressure  
 671 development, and its actual impact on inland Asian aridity. Indeed, the presence of loess-  
 672 like deposits since the Eocene, that were tentatively explained by the presence of an active  
 673 Siberian High at that time (Meijer et al., 2021; Licht et al., 2014), question its actual  
 674 dependence to the presence of Tian Shan and/or Mongolian orogens, which are usually dated  
 675 later in the Oligocene to late Miocene. In addition, the notion of threshold elevation is a key  
 676 point to tackle, as strongly non-linear responses of climate circulation with respect to this  
 677 parameter are expected. These considerations highlight the need for a finer understanding of  
 678 the actual drivers of today’s regional Asian monsoons. Given the demonstrated prevalence  
 679 of geography in shaping the Asian Monsoons, an accurate representation of its evolution is  
 680 therefore paramount to study their history over long time-scales.

## 681 4 Cenozoic paleogeography evolution

682 Simulations in this study have been performed with Eocene, Oligocene and Miocene  
 683 paleogeographic configurations from previous studies (Poblete et al., 2021; Tardif et al.,  
 684 2020; Sarr et al., 2022; Barbolini et al., 2020), which are coherent with geological history.  
 685 The Eocene and the early Miocene paleogeographic configurations are from Poblete et al.  
 686 (2021); Tardif et al. (2020) reconstruction at 40 and 20 Ma. The early Oligocene geography  
 687 is based on the Eocene geography with an homogeneous sea level drop of 70 meters to  
 688 account for sea level changes related to Antarctic ice-sheet building during the Eocene-  
 689 Oligocene transition (Miller et al., 2020). The mid and late Miocene geographies are based  
 690 on the PRISM4 (Dowsett et al., 2016) reconstruction used in PlioMIP2 (Haywood et al.,  
 691 2020) with modifications to include a southward shift of the Australian continent and an  
 692 emerged Sunda Shelf (see Sarr et al. (2022) for details). We in addition performed sensitivity  
 693 experiments to test the impact of specific topographic features on different characteristics  
 694 of the Asian monsoons.

695 In this section, we provide an overview of the main landmarks and persistent uncer-  
 696 tainties regarding the paleogeography evolution during the Cenozoic in our region of interest  
 697 (see Fig. 2). The paleogeographic configurations used in our simulations are introduced, for  
 698 each of these key regions (Fig. 3 and Table 1).

### 699 4.1 Asia

#### 700 4.1.1 Indo-Asia collision and land-sea distribution uncertainties

701 The initiation of the Indo-Asian collision, marking the closure of the Neotethys Ocean  
 702 (J. R. Ali & Aitchison, 2008; Chatterjee et al., 2013), is dated around  $\sim 50$  Ma (C. Wang  
 703 et al., 2014; D. J. J. van Hinsbergen et al., 2012; Dupont-Nivet et al., 2010; Jagoutz et  
 704 al., 2015; Lippert et al., 2014; W. Huang et al., 2015), or alternately around  $\sim 58$  Ma  
 705 (D. J. van Hinsbergen et al., 2019; Ingalls et al., 2016), via sedimentology and paleomagnetic  
 706 evidence. The morphology of the *Greater India* portion (i.e. northern India) before collision  
 707 remains unclear. Depending on the collision scenario considered (see Kapp and DeCelles

**Table 1.** List of reference simulations (in 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 explains the relationship between the different paleogeographic configurations, and therefore define the anomalies that can be tested (e.g. the effect of a lowered TP will be studied through the "LEo2\_TP - LEo2\_TPlow" difference, while the effect of  $p\text{CO}_2$  halving in the Eocene is obtained with the "LEo1\_REF - LEo1\_2X". Likewise, all paleogeography sensitivity experiment from the Middle to late Miocene will be compared to the "LMio\_smallT" simulation.

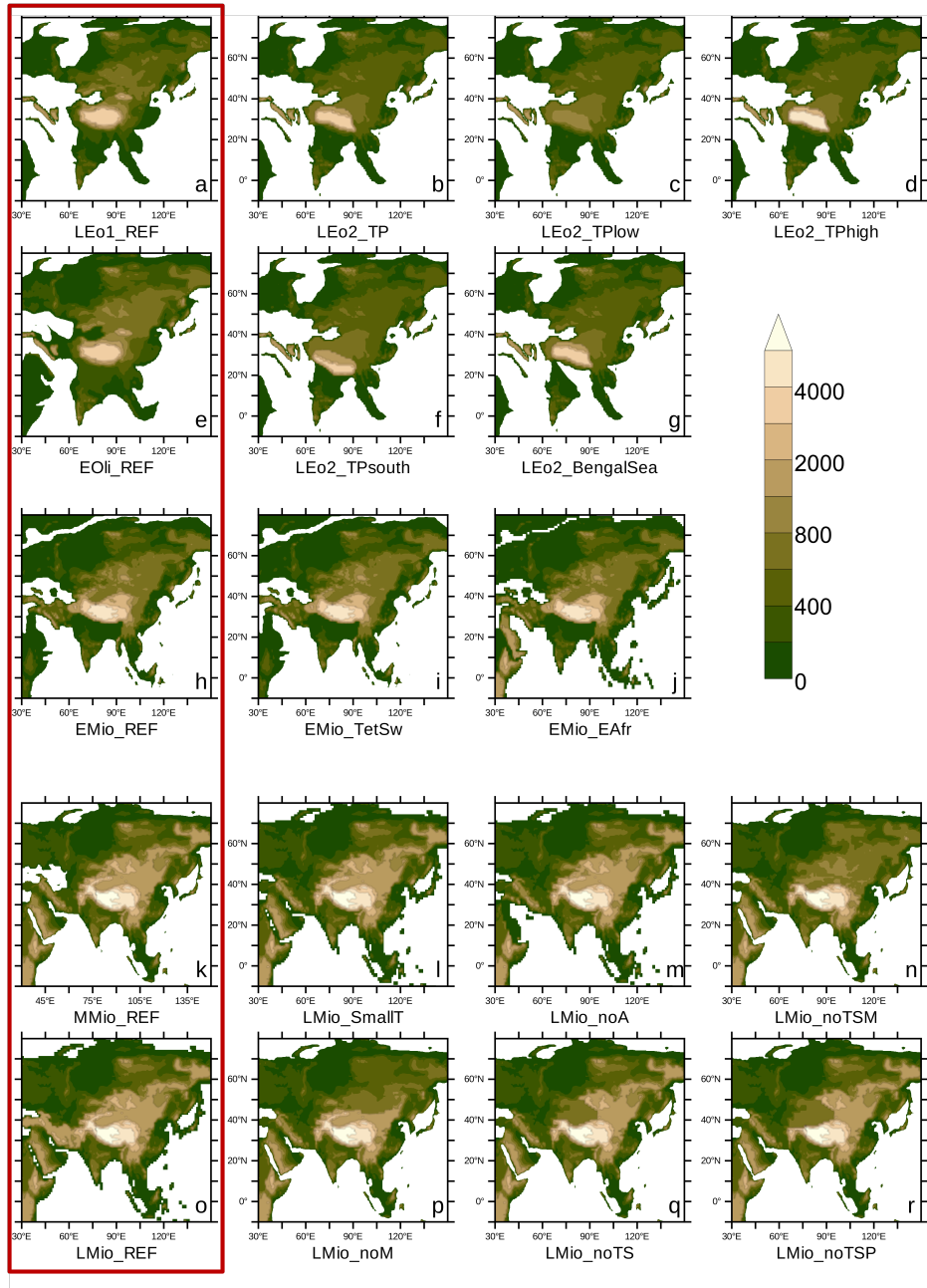
Simulation	$p\text{CO}_2$	Ice sheets	Paleogeography
<b>LEo1_REF</b>	<b>1120</b>	-	from (Tardif et al., 2020)
LEo2_TP	1120	-	from (Poblete et al., 2021)
LEo2_TPlow	1120	-	from LEo2_TP, with TP lowered to 800 m
LEo2_TPhigh	1120	-	from LEo2_TP, with TP raised to 4500 m
LEo2_TPsouth	1120	-	from LEo2_TP,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

(2019); Poblete et al. (2021) for a review), *Greater India* is represented as fully emerged (Ingalls et al., 2016; C. Wang et al., 2014), or partially flooded until the middle-late Eocene (D. J. J. van Hinsbergen et al., 2012; D. J. van Hinsbergen et al., 2019; W. Huang et al., 2015). From the Eocene onward, India's progressive migration to the North, indentation into the Asian continent and counterclockwise rotation (Molnar et al., 2010) triggered widespread orography changes in Asia, which are summarized below. The most salient of these events is of course the Tibetan Plateau and Himalayas uplift, and the spreading of deformation and uplift to peripheral terranes.

#### 4.1.2 Himalayas-Tibetan Plateau uplift history

The early uplift history of the Tibetan Plateau is highly debated. Oxygen isotopic paleoaltimetry studies suggest that the Gangdese (Southern TP, Lhasa terrane) and Qiangtang mountains (Central TP) are already as high as  $\sim 4000$ - $5000$  m during the early Eocene (Rowley & Currie, 2006; L. Ding et al., 2014; Xu et al., 2013; Xiong et al., 2020; C. Wang et al., 2014). The robustness of this method is however questioned by isotope-enabled climate models demonstrating that other factors than elevation affect the  $\delta^{18}\text{O}$  precipitation signature, such as the air masses provenance, climate change, or water recycling (Poulsen & Jeffery, 2011; Botsyun et al., 2016). Considering these additional factors, a revised Ti-





**Figure 3.** paleogeographic configurations used in this study. (a,e,h,k,o) Reference paleogeographic configurations for the late Eocene ( $\sim 40$  Ma), early Oligocene ( $\sim 34$  Ma), early ( $\sim 20$  Ma), mid ( $\sim 12$  Ma) and late Miocene ( $\sim 8$  Ma) are aligned in the first column (red box). Alternative configurations used for sensitivity tests are also presented, for the late Eocene (b,c,d,f,g), the early Miocene (i,j) and the mid to late Miocene (l,m,n,p,q,r)

betan Plateau paleo-elevation, most likely inferior to  $\sim 3000$  m in the Oligocene, is suggested (Botsyun et al., 2019). The presence of frost-intolerant fossil flora assemblage in Central TP in the middle Eocene (T. Su et al., 2020) and late Oligocene (T. Su, Farnsworth, et al., 2019), as well as clumped isotope measures (Xiong et al., 2022), advocating for a subtropical climate in this region, seems incompatible with a TP exceeding 4000 m. Therefore, a more complex morphology for this proto-TP, with the presence of a low-elevated ( $\sim 1500$  m) valley dividing the almost fully uplifted ( $\sim 3000$ - $4000$  m) Lhasa and Qiangtang terranes until the late Oligocene has been proposed (T. Su, Farnsworth, et al., 2019; Xiong et al., 2020; R. Spicer et al., 2020; Xiong et al., 2022).

In the periphery of the proto-TP, data collected from different basins offer contradictory information. In southeastern Tibet, at the extremity of the Qiangtang terrane, T. Su, Spicer, et al. (2019) use paleo-elevation calculated from the CLAMP method to suggest that the Markam Basin was already high in the late Eocene ( $\sim 3000$  m) and could have reached its modern elevation ( $\sim 3900$  m) by the earliest Oligocene. Alternately, paleo-elevation based on oxygen isotopic measurements of the nearby Jianchuan Basin (belonging to the Qiangtang terrane as well) could have not exceeded  $\sim 1200$  m in the late Eocene (Gourbet et al., 2017). The northern TP is proposed to remain relatively low (maximum  $\sim 2000$  m) in the Eocene before uplifting in the Oligocene or the early Miocene, based on  $\delta^{18}\text{O}$  measurements from the Hoh Xil Basin (Cyr et al., 2005; C. Wang et al., 2014). An early uplift of the northeastern TP is however suggested as early as the late Eocene by increased coniferous pollen proportion in the Xining Basin (Hoorn et al., 2012), although cooling and Paratethys Sea retreat were also shown to promote coniferous trees expansion in this region (Barbolini et al., 2020).

The history of the TP build-up from the Miocene onward is less controversial. The Central TP would have reached modern elevation by the early Miocene, while the northern TP uplift is completed by the late Miocene (C. Wang et al., 2014). Stratigraphy and detrital zircon analysis indicate that the western TP (Pamir-Karakoram-Hindu Kush) may have underwent several phases of uplift between the mid Eocene to latest Miocene (Bershaw et al., 2012; L. Li et al., 2021; Blayney et al., 2016; Carrapa et al., 2015). Finally, the Himalayas start uplifting in the early Miocene ( $\sim 20$  Ma) and reach their modern elevation by the mid-late Miocene ( $\sim 15$  Ma) (Gébelin et al., 2013; L. Ding et al., 2017; Xu et al., 2018; C. Wang et al., 2014; Webb et al., 2017).

#### 4.1.3 *Inland Asia and Myanmar mountain building*

Asian terranes located in the vicinity of the HTP complex are impacted by the Indo-Asian collision. In Myanmar, the Sino and Indo-Burman Ranges have distinct uplift histories. Indo-Burma ranges build-up in a three-step process in the late Eocene, at the Oligo-Miocene boundary and Pliocene respectively (Najman et al., 2020; Morley, 2018; Maurin & Rangin, 2009; Westerweel et al., 2020; Licht et al., 2018), as a result of the progressive indentation of the Burma block into the Eastern Himalayan collision zone. The Sino-Burman Ranges inherited old landforms from previous collisions between Indochina and Sibumasu terranes (Metcalf, 2013) and experienced major uplift in the mid-late Miocene ( $\sim 13$ - $9$  Ma) (Cook & Royden, 2008; Clark et al., 2005), as a result of the Tibetan crustal flow eastward propagation.

North of the Tibetan Plateau, the collision propagated through reactivated pre-existing tectonic structures and triggered the uplift of the inland Asia orogenic belt: the Tian Shan and the Mongolian (Altai, Hangay and Sayan) landforms (Jolivet et al., 2010). The southwestern and Central Tian Shan uplift initiation is very loosely constrained, either in the middle to late Miocene ( $\sim 15$  Ma, (Käbner et al., 2016)), late Oligocene to early Miocene ( $\sim 25$  Ma, (Bande et al., 2017; Macaulay et al., 2016, 2014)), or in the Eocene-Oligocene (D. Liu et al., 2017; Tapponnier & Molnar, 1979). The northeastern Tian Shan, together with the Sayan and Altai regions (in western Mongolia), likely remained quiescent until the

late Miocene (Bullen et al., 2003; Caves Rügenstein et al., 2017). After a global pulse of uplift at the latest Miocene ( $\sim 8$  Ma) for the whole Tian Shan, Sayan and Altai mountains (W. Yuan et al., 2006; Caves Rügenstein et al., 2017; Charreau et al., 2005), regional landforms would have reached their full elevation after a last Pliocene uplift phase ( $\sim 5$ -3 Ma, (Bullen et al., 2003; Caves Rügenstein et al., 2014; De Grave et al., 2007, 2009; Jolivet et al., 2007; Vassallo et al., 2007)). The Hangay Dome uplift (in central Mongolia) began in the mid-Oligocene (Caves Rügenstein et al., 2014; Cunningham, 2001), but the date of completion of the uplift remains debated, either in the mid-Miocene ( $\sim 13$  Ma, (Smith et al., 2016)), or the Pliocene ( $\sim 3$  Ma, (Yarmolyuk et al., 2008)).

#### 4.1.4 Evolution of Asia in our reference paleogeographic configurations and sensitivity experiments performed

Our late Eocene reference paleogeography (**LEo1\_REF**,  $\sim 40$  Ma, Fig. 3 a), previously introduced in (Tardif et al., 2020), displays a fully emerged *Greater India* and a TP elevation set to  $\sim 3000$  m. Considering the persistent controversy surrounding the extension of India and the TP morphology at that time, other configurations are tested to better understand the sensitivity of our results to these features. **LEo2\_TP** (Fig. 3 b) is an alternative late Eocene paleogeography, with a slightly different Tibetan Plateau shape (obtained with the "double collision model" in Poblete et al. (2021)), from which we derived sensitivity tests with a lowered Tibetan Plateau ( $\sim 800$  m, **LEo2\_TPlow**, Fig. 3 c) and a high Tibetan Plateau ( $\sim 4500$ , **LEo2\_TPhigh**, Fig. 3 d). Two additional tests are performed, with a Tibetan Plateau translated more to the South (**LEo2\_TPsouth**, Fig. 3 f), obtained with the collision model from Jagoutz et al. (2015) and the presence of a remaining Bengal Sea (**LEo2\_sea**, Fig. 3g), obtained with the collision model from D. J. J. van Hinsbergen et al. (2012); D. J. van Hinsbergen et al. (2019).

The early Oligocene simulation (**EOLI\_REF**,  $\sim 33$  Ma, previously introduced in Barbolini et al. (2020)), is designed to represent the Asian climatic state right after the Eocene-Oligocene Transition (Fig. 3 e). We use the same late Eocene paleogeography from **LEo1\_REF**, in which the sea level is lowered by 70 m to account for an ice sheet over Antarctica with a modern extent (it is an end-member scenario). This leads to important changes in terms of land-sea distribution, principally in the Paratethys Sea region, detailed in section 4.2.3.

The early Miocene paleogeography (**EMio\_REF**,  $\sim 20$  Ma, Fig. 3 h) (previously introduced in Sarr et al. (2022) and Burls et al. (2021), from Poblete et al. (2021)) displays an Indian subcontinent translated  $5^\circ$  further north compared to the Eocene, and an uplifted Central TP close to its present elevation ( $\sim 5000$  m), while western and northeastern TP portions are lower ( $\sim 2000$ -3000 m). In the middle (**MMio\_REF**,  $\sim 12$  Ma, Fig. 3 k) and late Miocene (**LMio\_REF**,  $\sim 8$  Ma, Fig. 3 o) paleogeographic configurations, the Himalayas and Tibetan Plateau have reached their present day configuration, along with the peripheral landforms in Myanmar and inland Asia (Fig. 3). Sensitivity tests are performed on inland Asian landforms, lowering to  $\sim 800$  m the elevation of Mongolia (**LMio\_noM**, Fig. 3 p), the Tian Shan (**LMio\_noTS**, Fig. 3 q), both Mongolia and Tian Shan (**LMio\_noTSM**, Fig. 3 n), and finally, the Tian Shan and Pamir (**LMio\_noTSP**, Fig. 3 r).

## 4.2 The Peri-Tethys region evolution under the complex interplay of tectonics and global sea level changes

Throughout the Cenozoic, global sea level fluctuations in an active tectonic convergence context have led to a profound paleogeographic reorganization and a general increase in continentality of the regions encompassing the Mediterranean Sea, Middle East and inland Asia. We resume here the main paleogeography changes (in terms of sea connections between Indian, Atlantic and Arctic Oceans).

#### 825 **4.2.1 Retreat of the Paratethys Sea and closure of the Neotethys Ocean**

826 During the Paleocene-Eocene, Eurasia is largely flooded by the proto-Paratethys, an  
 827 epicontinental sea connected to the Arctic Ocean *via* the Turgai Strait and to the Neotethys  
 828 Ocean (Meulenkamp & Sissingh, 2003; Golonka, 2009). Paleoenvironmental analysis suggest  
 829 a maximal extension of the proto-Paratethys Sea reaching the Tajik and Tarim Basins, in  
 830 inland Asia, at the Paleocene-Eocene transition (~56 Ma), followed by a long-term westward  
 831 retreat, punctuated by marine transgressions of decreasing magnitude (Kaya et al., 2019;  
 832 Bosboom et al., 2014b, 2017; Carrapa et al., 2015). In Central Asia, the shrinkage of the  
 833 proto-Paratethys Sea during the Eocene is attributed to far-field effect of the Indo-Asia  
 834 collision and to the early indentation of the Pamir and TP, leading to the progressive basins  
 835 infilling with sediments (Kaya et al., 2019; Carrapa et al., 2015). The build-up of the  
 836 Antarctic ice-sheet initiated in the late Eocene, resulted in a sea-level drop that culminated  
 837 at the EOT (~33.5 Ma, sea-level fall estimated at ~70 m by Miller et al. (2020)), which  
 838 drastically reduced the extent of the Paratethys and favored its isolation (Kaya et al., 2019;  
 839 Bosboom et al., 2014b, 2017). Water exchanges between the Paratethys and global oceans  
 840 are progressively restricted, as attested by  $\delta^{18}\text{O}$  and  $\delta^{13}\text{C}$  of foraminifera shells diverging  
 841 increasingly from global oceanic signals at that time (Ozsvárt et al., 2016). The Paratethys  
 842 Sea is fully isolated from the Mediterranean Sea at ~12 Ma, with the final uplift of the  
 843 Dinarides belt (Harzhauser & Piller, 2007). A connection with the Mediterranean Sea is  
 844 restored between 6.9–6.1 Ma when the Aegean region subsides during the latest Miocene  
 845 (Krijgsman et al., 2020).

846 The Tethyan Seaway (or Mesopotamian Seaway) which connected Indian and Atlantic  
 847 Oceans remained fully open before the Miocene (Straume et al., 2020).  $\epsilon\text{Nd}$  data suggest  
 848 that it became strongly restricted around ~22 Ma (Bialik et al., 2019), with intermittent  
 849 periods of closure, as attested by evidence of mammal exchanges between Eurasia and  
 850 Africa through the Gomphotherium landbridge (Rögl, 1999; Harzhauser & Piller, 2007).  
 851 The Tethyan Seaway permanently closed at ~14 Ma (Bialik et al., 2019), due to combined  
 852 effect of Arabian-Eurasian plates collision and glacio-eustatic sea level fall associated with  
 853 the cooling following the Mid-Miocene Climatic Optimum (MMCO).

#### 854 **4.2.2 Middle-East: Eastern Anatolian-Iranian Plateau and Zagros Moun-** 855 **tains**

856 The Middle-East experienced multiple changes in land-sea distribution and topogra-  
 857 phy, forced by the collision of the Arabian plate with Eurasia, which drove the uplift of  
 858 the Zagros mountains and of the Iranian Plateau. The exact timing of the Arabia-Eurasia  
 859 collision in eastern Anatolia remains highly debated with estimated date ranging from the  
 860 Eocene-Oligocene (Karaođlan et al., 2016; Darin et al., 2018; McQuarrie & Hinsbergen,  
 861 2013; Pirouz et al., 2017), the early Miocene (Okay et al., 2010; Gülyüz et al., 2020), to  
 862 the late Miocene (H. Su & Zhou, 2020; Z. Zhang et al., 2017). Most studies however pro-  
 863 pose an early Miocene age (~20 Ma) for the hard collision related to the arrival of thick  
 864 Arabian crust along the Bitilis suture zone (Okay et al., 2010; Cavazza et al., 2018; Gülyüz  
 865 et al., 2020). After the collision, the Zagros orogen built up during three successive pulses  
 866 occurring within the last ~20 Ma (Agard et al., 2011; Mouthereau, 2011).

867 The Alborz mountains uplift (northwest of the Zagros) is dated during the middle  
 868 Miocene (~17.5–13 Ma), according to sediments stable isotopic signatures (Ballato et al.,  
 869 2010, 2015) and may have continued until the latest Miocene (Mouthereau et al., 2012).  
 870 A Miocene age for the commencement of the eastern Anatolian Plateau growth has been  
 871 estimated based on the youngest marine unit of the eastern Anatolian Plateau (~17 Ma)  
 872 (Gülyüz et al., 2020) and apatite fission-track data (18–13 Ma) (Okay et al., 2010; Karaođlan  
 873 et al., 2016). Stratigraphic evidence suggest a rise due to crustal shortening and thickening  
 874 occurring between 15 and 12 Ma (Mouthereau et al., 2012). Numerical simulations suggest  
 875 a buildup of the Plateau within ~15–20 Ma after the continental collision, likely sustained

876 by slab break-off, mantle flow disruption and associated changes in dynamic topography,  
 877 followed by isostatic adjustment (François et al., 2014). With this mechanism, the modern  
 878 elevation ( $\sim 1500$  m.a.s.l) is only reached during the late Miocene ( $\sim 7-10$  Ma).

#### 879 **4.2.3 Evolution of the Peri-Tethys region in our reference paleogeographic** 880 **configurations and sensitivity experiments performed**

881 Our paleogeographic reference configurations are consistent with the geological history  
 882 of the Paratethys Sea demise and with uplift phases in the region. In our late Eocene  
 883 geography (**LEo1\_REF**, Fig. 3 a) the proto-Paratethys Sea fills the Tarim Basin ; it is still  
 884 connected to the Neotethys and Indian Oceans, but not to the Arctic, as the Turgai Strait  
 885 is already closed. In the Oligocene reconstruction (**EOLI\_REF**, Fig. 3 e), the 70 m sea  
 886 level drop (Miller et al., 2020) leads to the emergence of continental portions in Arabia and  
 887 North-Africa, and to the drying of the Tarim Basin and lowlands north of the Turgai Strait,  
 888 which all conspire to increase greatly Asian continentality. The former shallow connection  
 889 between the Paratethys and the Tethys ceases in the Pamir-Zagros region at that time. In  
 890 the early Miocene geography (**EMio\_REF**, Fig. 3 h), an emerged land-bridge interrupts  
 891 the Tethyan seaway that previously connected the Mediterranean Sea and the Indian Ocean,  
 892 but shallow sea persists at the location of the present-day Arabian Peninsula. Given the  
 893 uncertainty surrounding the precise closure time of the seaway, an alternative early Miocene  
 894 configuration is also tested, with a 120 m deep Tethyan seaway (**EMio\_TetSw**, Fig. 3 i).  
 895 In the middle Miocene reference simulation (**MMio\_REF**, Fig. 3 k), the Iranian Plateau  
 896 and Zagros mountains are halfway uplifted ( $\sim 1000$  m) compared to present-day and the  
 897 Paratethys extends up to  $60^\circ\text{E}$ . By the late Miocene (**LMio\_REF**, Fig. 3 o), the Paratethys  
 898 is strongly reduced and the Iranian landforms have reached their modern elevation. An  
 899 intermediate reconstruction is proposed, testing the impact of the Paratethys retreat only,  
 900 in the mid-late Miocene (**LMio\_smallT**, Fig. 3 l). Note that LMio\_smallT is the mid-late  
 901 Miocene paleogeography to which all others late and middle Miocene simulations can be  
 902 compared to, to single out the effect of regional changes.

### 903 **4.3 East Africa and Arabian Peninsula evolution**

904 While much attention has been paid to the Eastern African topography evolution in the  
 905 Neogene due to its implication on early hominins development, its history fits in a broader  
 906 paleogeographic context rooted in the late Eocene to early Oligocene times (see Couvreur  
 907 et al. (2021); Guillocheau et al. (2018) for a synthesis). Landforms in East Africa (Kenyan  
 908 and Ethiopian plateaux) developed during the late Cenozoic, triggered by a combination  
 909 of tectonics, basaltic flooding, and large wave-length deformation related to the African  
 910 superswell activity (Moucha & Forte, 2011; Faccenna et al., 2019; Roberts et al., 2012).  
 911 Uplift in the Ethiopian region would have began during the Eocene, with doming event  
 912 sustained by dynamic topography (Roberts et al., 2012; Faccenna et al., 2019) that then  
 913 propagates to Kenya during the middle Miocene. In the early Oligocene, basaltic flooding  
 914 in Ethiopia would have contributed to the formation of these large-scale elevated features  
 915 (Sembroni et al., 2016). Rifting initiated during the middle Miocene, with main uplift of  
 916 rift shoulders (Wichura et al., 2015). Overall, past elevations in the course of the Cenozoic  
 917 are poorly constrained in East Africa, but modern elevations were most likely reached by  
 918 the late Miocene to early Pliocene (Couvreur et al., 2021).

#### 919 **4.3.1 Arabian Peninsula**

920 The Arabian Peninsula records successive marine and freshwater environments from  
 921 the early Oligocene to the late Miocene, as attested by paleo-ichtthyofaunas (Otero &  
 922 Gayet, 2001). Before the Neogene, the Arabian plate is assumed to be flat and low ele-  
 923 vated (Daradich et al., 2003), making it particularly sensitive to sea-level changes. It is  
 924 therefore largely submerged during the Eocene, and its southwestern part becomes par-



925 tially emerged after the sea-level fall at EOT (Barrier et al., 2018). During the Miocene,  
 926 the Arabian plate is tilted when transiting over the East-Africa mantle plume (Vicente de  
 927 Gouveia et al., 2018), leading to the uplift of the Red sea margin, and the flooding of its  
 928 northeastern corner, due to mantle convective drawdown (Daradich et al., 2003; Moucha  
 929 & Forte, 2011). This mechanism favored temporary transgression phases in the subdued  
 930 northeastern region in early Miocene times (Barrier et al., 2018), until its full emergence in  
 931 the late Miocene, possibly favored by global sea-level fall after the MMCT (Golonka, 2009;  
 932 Harzhauser & Piller, 2007; Miller et al., 2020).

#### 933 *4.3.2 Evolution of East Africa and Arabia in our reference paleogeographic* 934 *configurations and sensitivity experiments*

935 Our reference paleogeographic configurations (Fig. 3 a,e,h,k,o) display a low East  
 936 Africa during the Eocene-Oligocene, which rises to  $\sim 750$ - $1,000$  m in the early Miocene and  
 937 a modern-like elevation in the mid and late Miocene reconstructions. Given the large uncer-  
 938 tainties on east African topography precise elevation, we perform a sensitivity experiment on  
 939 an early Miocene paleogeography, with fully uplifted east African landforms (**EMio\_EAfr**,  
 940 Fig. 3 j). Additionally, a mid-late Miocene configuration displaying an almost fully flooded  
 941 Arabian Peninsula is tested (**LMio\_noAr**, Fig. 3 m), to account for its high sensitivity to  
 942 sea level fluctuations at that period.

## 943 **5 Materials and Methods**

### 944 **5.1 IPSL-CM5A2 model**

945 We performed the simulations with the IPSL-CM5A2 Earth System Model (Sepulchre  
 946 et al., 2020), an updated version of the IPSL-CM5A model (Dufresne et al., 2013) suitable  
 947 for deep-time paleoclimate simulations. It is composed of the atmospheric model LMDz5A  
 948 (Hourdin et al., 2013), the land surface and vegetation model ORCHIDEE (Krinner et al.,  
 949 2005) and the oceanic model NEMO (v3.6) (Madec, 2016) that also includes the LIM2  
 950 model (Fichefet & Maqueda, 1997) for sea-ice. OASIS coupler ensures synchronization  
 951 between the different model components (Valcke et al., 2006). Atmospheric component  
 952 has a nominal horizontal resolution of  $96 \times 95$  grid points ( $3.75^\circ$  in longitude by  $1.9^\circ$   
 953 in latitude) with 39 irregularly distributed vertical levels from the surface to 40 km high.  
 954 The land surface ORCHIDEE model is coupled with LMDZ5a and redirects runoff water  
 955 toward the ocean. The vegetation cover is represented through 11 Plant Functional Types  
 956 (PFTs), including one describing bare soil. NEMO uses a tripolar grid, with two poles in  
 957 the Northern hemisphere to avoid singularity (Madec & Imbard, 1996). It has a resolution  
 958 of  $2^\circ$  by  $2^\circ$  refined at  $0.5^\circ$  in the equatorial region, with 31 vertical levels, whose thickness  
 959 span from 10 meters near the surface to 500 meters in the deep ocean. Full description of  
 960 IPSL-CM5A2 can be found in Sepulchre et al. (2020); Dufresne et al. (2013).

961 IPSL-CM5A2 has been previously used for paleoclimatic simulations of the Miocene  
 962 (Burls et al., 2021; Sarr et al., 2022) and the Eocene (Toumoulin et al., 2020; Tardif  
 963 et al., 2020; Barbolini et al., 2020; Tardif et al., 2021; Toumoulin et al., 2022). The ability  
 964 of the model to simulate the modern Asian climate was previously assessed (Sepulchre et  
 965 al., 2020; Tardif et al., 2020) and a comparison with observations is provided (SI, Fig. 15).  
 966 While general atmospheric circulation patterns are well reproduced, IPSL-CM5A2 overall  
 967 underestimates precipitation amounts and delays inland monsoonal precipitation onset by  
 968 about a month, a feature shared with IPSL-CM5A version (Sepulchre et al., 2020; Valdes  
 969 et al., 2017).

### 970 **5.2 Experiments design**

971 The boundary conditions are summarized in Table 1. Experiments at Late Eocene  
 972 are considered ice-free, as  $p\text{CO}_2$  is above the threshold for permanent Antarctic glaciation

973 (Ladant et al., 2014; DeConto & Pollard, 2003). Unipolar glaciation, with Antarctic ice-  
 974 sheet only, is used for the Oligocene and early Miocene and both Greenland and Antarctic  
 975 ice-sheets are accounted for in the late Miocene (Bierman et al., 2016) configurations. CO<sub>2</sub>  
 976 concentration varies from 1,120 ppm for the late Eocene to 560 ppm in the late Miocene  
 977 (Table 1) Foster et al. (2017); Rae et al. (2021), while other greenhouse gases are kept at  
 978 preindustrial values. Given the persistent uncertainties regarding precise  $p\text{CO}_2$  concentra-  
 979 tion during the Cenozoic, we performed two additional sensitivity tests on this parameter,  
 980 in the late Eocene (**LEo1\_2X**, 560 ppm), and in the late Miocene (**LMio\_1.5X**, 420 ppm).  
 981 The solar constant was adjusted according to Gough (1981) and set to 1360.19 W.m<sup>-2</sup> for  
 982 late Eocene and early Oligocene simulations, to 1362.92 W.m<sup>-2</sup> for early Miocene simu-  
 983 lations and to 1364.30 W.m<sup>-2</sup> for late and middle Miocene simulations. Although orbital  
 984 variations are known to impact monsoonal circulation (Zhisheng et al., 2015; P. Wang et  
 985 al., 2005; Tardif et al., 2021), we prescribed a modern-like configuration to focus on the  
 986 impact of paleogeography on the monsoon system. Considering the lack of congruent global  
 987 vegetation reconstruction over the Cenozoic, the same PFT map was used throughout the  
 988 experiments, prescribing latitudinal bands of idealized vegetation cover based on modern  
 989 distribution, as done in previous studies (Tardif et al., 2020; Laugié et al., 2020). The  
 990 reference experiments have been run for 3000 years and deep ocean layers reached quasi-  
 991 equilibrium after ~2000 years of integration. Sensitivity experiments were restarted from  
 992 there reference experiments an run for 300 to 500 years until reaching stable sea surface  
 993 temperatures. Model outputs are averaged over the last 50 years.

### 994 5.3 Monsoon metrics

995 The evolution of Asian monsoons back in time is assessed through specific markers  
 996 that are characteristic of the present-day phenomenon. We favored SAM and EAM features  
 997 offering means of comparison with paleoclimate records, which typically involve precipitation  
 998 amounts, seasonality and wind direction and strength. Several features of the monsoons,  
 999 however, are not recorded by proxies, or in a very indirect way. This is for example the case  
 1000 of the pressure patterns, and of all mid-to-high tropospheric features associated with the  
 1001 modern monsoons. In that aspect, numerical simulations are a valuable tool to grasp the  
 1002 full signature of eventual paleo-monsoons.

1003 For both SAM and EAM, we evaluated the evolution of precipitation seasonality using  
 1004 the Monsoon Precipitation Index (MPI) (B. Wang & Ding, 2008), expressed as follow:

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

1005 where the seasonal range of precipitation is the difference between May to September  
 1006 (MJJAS) precipitation minus November to March (NDJFM) precipitation. Regional oc-  
 1007 currence of a monsoon-like seasonality is considered providing a seasonal range superior to  
 1008 300 mm and a majority of precipitation falling during the extended summer season (MPI  
 1009 > 0.5). This criteria has the benefit of not being exclusive (with or without monsoon), as  
 1010 it proposes a discrete metric for seasonality amplitude (from 0 to 1). It also imposes the  
 1011 wet season to be in summer, as in the present-day monsoon region, as opposed to other  
 1012 existing precipitation metrics (such as the ratio of the 3 wettest month over the 3 driest  
 1013 month precipitation, commonly used in paleobotanic studies), which enables us to test the  
 1014 presence of modern-like climate patterns.

1015 The wind patterns, described through their shape and strength, are the second funda-  
 1016 mental characteristic observed. The latitudinal migration of the Somali Jet is tracked over  
 1017 the Arabian sea (Fluteau et al., 1999). The wind shear between low and high troposphere  
 1018 over India, that describes the vigor of the SASM zonal circulation and its inclusion in the  
 1019 Walker circulation, is calculated through the Webster-Yang Index ( $U_{850}-U_{200}$ , calculated in  
 1020 an area from 0°:20°N to 40°E:110°E) (Webster et al., 1998). The positive thermal anomaly

1021 observed at 200-300 hPa over northern India in summer has additionally been used as a  
 1022 marker of the deep convection characterizing the modern SASM (Boos & Kuang, 2010;  
 1023 Acosta & Huber, 2020). This feature is due to the important latent heat release provoked  
 1024 by the condensation of moisture in convecting air masses and in air masses ascending on the  
 1025 southern flank of the Himalayas, which heat the high troposphere (see Fig. 14 c). Fewer  
 1026 criteria exist to formally describe the EAM. In addition to the precipitation seasonality de-  
 1027 scribed through the Monsoon Precipitation Index, we tracked the EASM via the description  
 1028 of the latitudinal displacement of the Jet Stream with respect to Tibetan and inland Asia  
 1029 landforms, and the surface wind patterns that constitute the Meiyu-Bayu front at present-  
 1030 day. The EAWM is tracked principally through the evolution of temperature and pressure  
 1031 gradients, leading to the Siberian High formation in winter, and of its associated wind pat-  
 1032 terns. Here we used the EAWM index defined by L. Wang and Chen (2014) that accounts for  
 1033 both the North-South (Siberia-Maritime continent) and West-East (Siberia-North Pacific)  
 1034 pressure gradients and is expressed as follow:

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

1035 where  $\text{SLP}_{\text{S}}$ ,  $\text{SLP}_{\text{NP}}$  and  $\text{SLP}_{\text{MC}}$  is the area-averaged Sea Level Pressure (SLP) over  
 1036 Siberia (40°:60°N, 70°:120°E), North Pacific (30°:50°N, 140°E:170°W) and the Maritime con-  
 1037 tinent (20°S:10°N, 110°:160°E), respectively.

## 1038 6 Results

1039 We present the results obtained for the five reference simulations as well as for the  
 1040 different sensitivity numerical experiments, first for the South Asian Monsoon, then for the  
 1041 East Asian Monsoon domain. Main outcomes in terms of seasonal winds and precipitations  
 1042 are qualitatively summed up in Table 6.

### 1043 6.1 South Asian summer monsoon controlled by the orography surround- 1044 ing the Indian Ocean

#### 1045 6.1.1 *Spatial and seasonal precipitation patterns evolution*

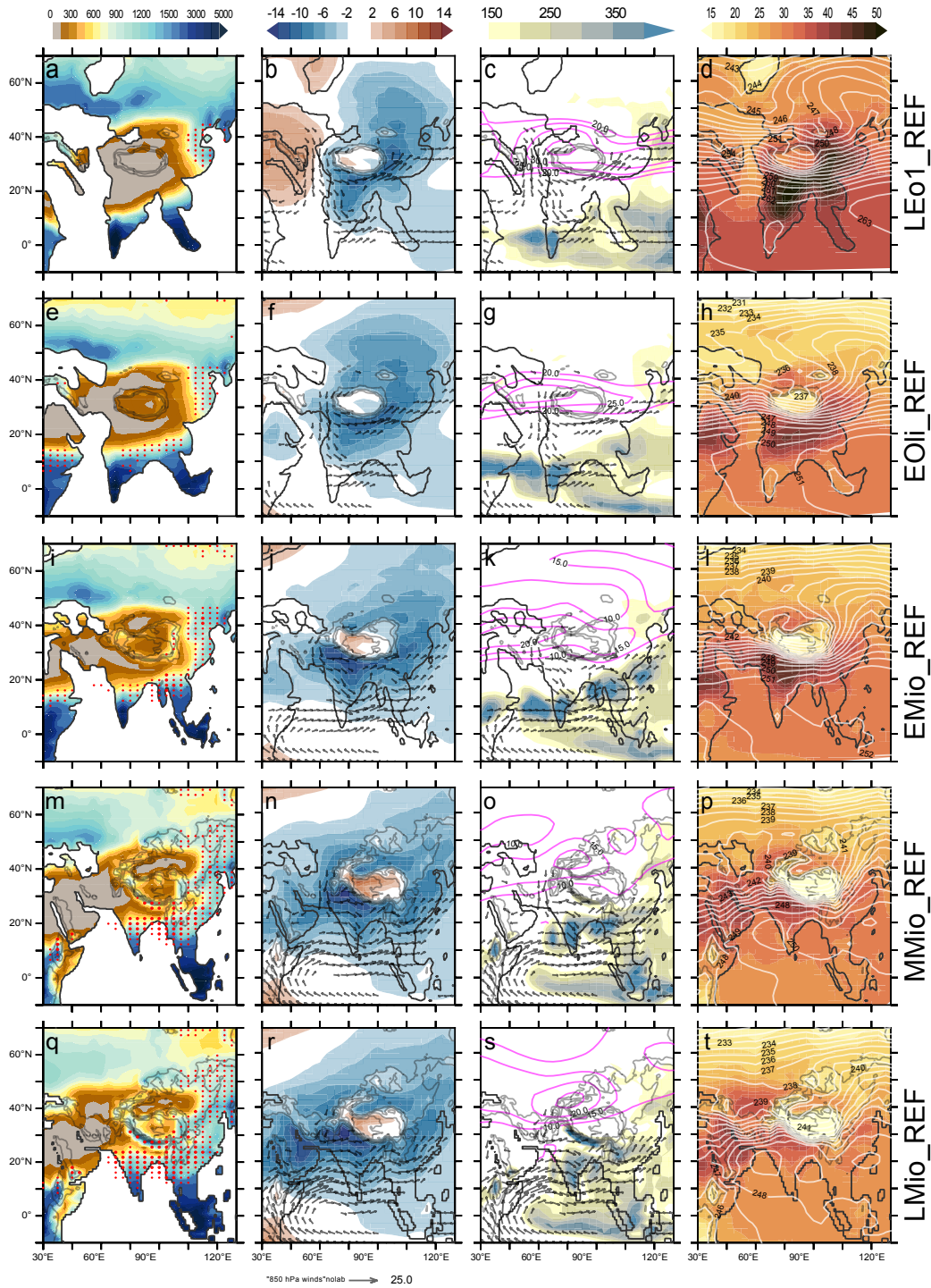
1046 In the late Eocene a wide arid region ( $< 500$  mm/yr) spreads from northern India  
 1047 to north of the proto-Tibet (Fig. 4 a). High mean annual precipitation amounts ( $> 3000$   
 1048 mm/yr) are simulated over equatorial regions, in southern India and southeastern Asia.  
 1049 There is no monsoon-like seasonality according to the MPI index in India, because the  
 1050 rainy season in this region occurs in winter (as previously described in Tardif et al. (2020)).  
 1051 The precipitation pattern evolves from the early Oligocene onward, with the progressive  
 1052 translation of the arid region to Arabia and Northern Africa (Fig. 4 e,i,m,q). This in  
 1053 turn allows the penetration of rainfall in summer in South-Eastern Asia, and the onset  
 1054 of a monsoon-like seasonality in this region, as well as a reinforcement of monsoon-like  
 1055 seasonality in Eastern Asia.

1056 In the course of the Miocene, as India drifts northward, its southern tip receives less  
 1057 precipitation ( $\sim 1500$  mm/yr in the late Miocene, Fig. 4 q), and the intensity and extent of  
 1058 highly seasonal precipitation increases greatly over southern and eastern Asia (red dotted  
 1059 regions in Fig. 4 e,i,m,q). Orographic precipitation over the southern flank of the Himalayas  
 1060 initiate by the mid Miocene simulation ( $\sim 900$  mm/yr, Fig. 4 m), and further increase in the  
 1061 late Miocene ( $\sim 1200 - 1500$  mm/yr, Fig. 4 q). These changes in precipitation patterns in  
 1062 both space and time are triggered by a profound reorganization of surface pressure patterns  
 1063 and winds throughout the Cenozoic.



Overview of the effect simulated in each region. Abbreviations stand for Transition Area (TA), cardinal directions (E, W, S, N), Basins mentioned in Fig. 1 (eg. Tadjik E)

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



**Figure 4.** Evolution of the 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 2m (shading, Celsius) and 300 hPa atmospheric temperature (white contour, Kelvin). On all captions, topography is overlain in gray contour each 1000 m.

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### 6.1.2 Summer pressure and wind patterns reorganization

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The summer (June to August, JJA) sea level pressure (SLP) patterns in Eurasia in the late Eocene experiment (Fig. 4 b) contrast strongly with the modern ones (Fig. 14 d). While a low-pressure center driven by extreme summer temperature forms over southeastern Asia (up to 50°C over *Greater India*, Fig. 4 d), the Paratethys Sea and Neotethys Ocean are the locus of a wide high pressure cell (Fig. 4 b). This induces low tropospheric anticyclonic wind patterns, that counteract potential moisture advection from the Indian Ocean towards South Asian regions. From the early Oligocene onward, the Paratethys Sea and Neotethys shrinkage increase the Eurasian continentality and emerging landforms, such as Anatolia and Arabia, become the locus of high summer temperature (35-45°C, Fig. 4 l,p,t). This leads 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 only attains its current values (~14 hPa over Arabia and northern India) by the late Miocene (Fig. 4 r).

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Mirroring this progressive strengthening and widening of the summer low-pressure belt over South Asia and Arabia, humidity-loaded air masses are gradually advected from the Indian Ocean towards the continent over the Cenozoic. A weak cross equatorial flow (<4 m/s against ~25 m/s today), confined to low latitudes (<10°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 proto-Somali Jet, mostly zonal and confined to low latitudes (<11°N) is simulated in the early Miocene simulation (Fig. 5 b, simulation EMio\_REF). From the mid Miocene onward, 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 and Fig. 5 b, simulations MMio\_REF and LMio\_REF).

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The simulated tropospheric circulation above southern Asia deeply evolves since the Late Eocene, as exemplified by the calculation of 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 through the Miocene (from ~16 to 21 m/s, Fig. 5 d, simulations EMio\_REF, MMio\_REF and LMio\_REF). Finally, while heating of the high troposphere (~300 hPa) due to deep convection is observed in most experiments (Fig. 4 d,h,l,p,t), its settlement over Northern India is only reached by the mid Miocene experiment, following the increase in summer convective precipitation in this region driven by orographic lifting.

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### 6.1.3 Limited effect of $p\text{CO}_2$ drawdown and Tibetan Plateau changes on South Asian summer monsoon evolution

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The non-linearity in the response of the Somali Jet through time calls for a more thorough analysis of the different forcing at stakes between each of these simulations. In this section, we take advantage of our different alternative paleogeography reconstructions and  $p\text{CO}_2$  to attribute the simulated changes to their respective underlying drivers. Overall, the SASM presents a low sensitivity to  $p\text{CO}_2$  changes, when compared to paleogeographic forcings. This is true for the late Eocene experiments, where a change from 1120 ppm to 560 ppm only induces minimal variations in the WYI (Webster-Yang Index) strength, summer northward ITCZ migration (difference between experiments LEo1.2X and LEo1\_REF in Fig. 5 a,c), or monsoon-like seasonality extension (Fig. 7 f). This is also the case for the late Miocene experiments, with a switch from 560 ppm to 420 ppm (difference between experiments LMio.1.5X and LMio\_REF in Fig. 5 b,d).

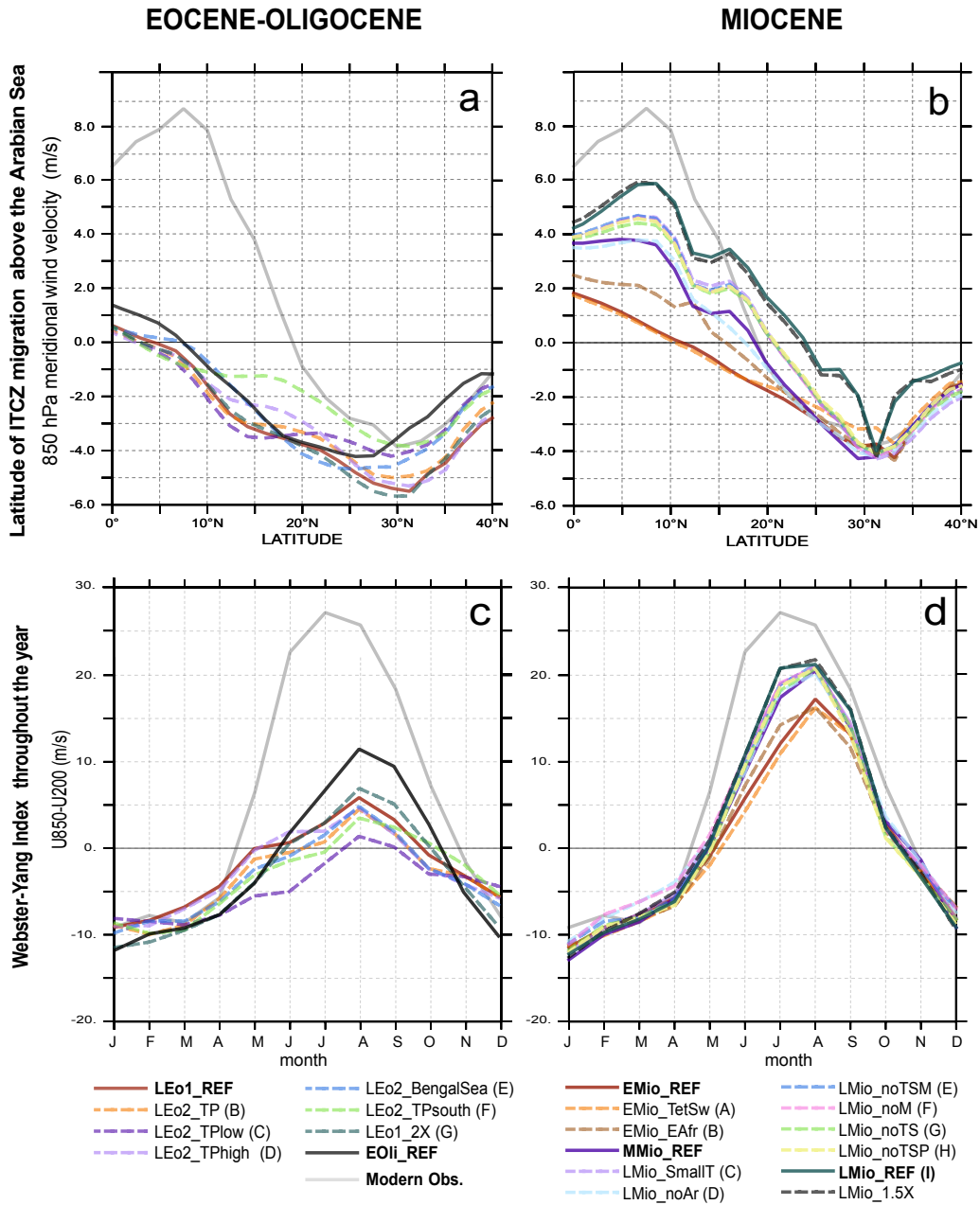
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The paleogeography of *Greater India* and Tibetan regions have been recurrently proposed as a driver of monsoon onset or intensification (R. Zhang et al., 2012; Yu et al., 2018; Molnar et al., 1993). While alternative TP configuration lead to slightly different results (detailed below), their impact on the SASM large-scale wind circulation remains limited in



**Figure 5.** Estimation of the South Asian monsoon strength evolution 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 (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 full lines, while sensitivity tests are in dotted lines.

our simulations (Fig. 5 a,c and Fig. 6). In the late Eocene, increasing TP elevation strengthens the zonal wind shear over India (3000 m, WYI =  $\sim$ 6 m/s, for experiment LEO2\_TP, Fig. 5 c), when compared to Eocene experiment with low topography in the region (800 m, WYI = 1 m/s, LEO2\_TPlow). Additionally, the magnitude of the WYI gradually increases with the TP latitude, with values spanning from 3 m/s when its southern edge is at  $\sim$ 20°N (LEO2\_TPsouth), 4 m/s (southern edge at  $\sim$ 25°N in LEO2\_TP), to finally 6 m/s (southern edge at  $\sim$ 30°N, experiment LEO1\_REF). The sea inlet that flooded the northern *Greater India* during the late Eocene, as sometimes suggested in the literature (D. J. J. van Hinsbergen et al., 2012; D. J. van Hinsbergen et al., 2019), limits the northward migration of the ITCZ in summer (from  $\sim$ 7 to 3°N between experiments LEO2\_BengalSea and LEO2\_TP, Fig. 5 c). It also produce a decrease in year-round precipitation, especially in South and inland Asia (Fig. 6 m,n,o).

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

The changes induced by the Tibetan plateau uplift or by its latitudinal location on the South Asian climate are however out-competed by far by the effects of other paleogeography changes, namely a global sea-level fall, reflecting the Antarctica glaciation at the EOT and the uplift of peripheral terranes (Fig. 5). Indeed, the increase in continentality driven by the sea level fall, together with the  $p\text{CO}_2$  halving, induces a colder and drier winter (Fig. 7 g) as well as hotter and wetter summer (Fig. 7 h), which results into a monsoon-like seasonality signal stretching over Southeastern Asia (Fig. 7 i). It also translates into a doubling of the WYI (from 6 to 11 m/s between experiments EOLI\_REF and LEO1\_REF, Fig. 5 c) and a strong reorganization of summer Jet Stream behavior (see dedicated section below).

Likewise, increasing continentality during the Miocene appears to be the key leverage for inducing a more pronounced migration of the ITCZ over the Arabian Sea (from 18 to 21°N in more continental case, Fig. 5 b), as well as a slight increase of the WYI (Fig. 5 d). This is achieved either with the emergence of the Arabian platform (Fig. 8 g,h,i), or with the narrowing of the Neotethys Ocean and the shrinkage of the Paratethys Sea (Fig. 8 d,e,f). In both cases, the amplification of the summer low pressure belt results into increased moisture advection from the Arabian Sea, and therefore increased summer precipitation in SAM and EAM domains. The overall precipitation seasonality increase is however not homogeneous, because of contrasted effects on the winter season precipitation. Our results point that the Arabian platform emergence, by amplifying the winter Siberian high pressure would dry Eastern Asia and southeastern Tibet, which would increase precipitation seasonality in parts of SAM but mostly in EAM domain (Fig. 8 d,f). The retreat of the Paratethys Sea, on the other hand, would mostly increase precipitation seasonality in the South Asian monsoon domain (Fig. 8 i). Moreover, a clear dichotomy emerges between Central Asia, that becomes drier after the Paratethys Sea retreat, and all surrounding regions that become wetter (Fig. 8 i).

Additionally, the successive buildup of both East African (difference between experiments EMio\_EAfr-EMio\_REF) and Iran-Zagros landforms (difference between experiments LMio\_REF-LMio\_smallT) is critical for the amplification of the ITCZ migration and strengthening of SAM. Each uplift triggers a  $\sim$ 4-5° northward migration of the ITCZ over the Arabian Sea in summer, as well as a slight increase in July WYI (Fig. 5 b,d). The uplift of East African landforms under Early Miocene conditions reinforces the Somali Jet and brings moisture to Iran-Zagros and SAM domain in summer (Fig. 9 e), which results in intensification of monsoon-like seasonality on the Himalayan foothills and in Southeastern Asia (mainly Myanmar, Fig. 9 f). Iran-Zagros landforms final uplift under late Miocene conditions promotes winter aridity and summer precipitation in SAM domain through a strong strengthening of the Somali Jet, resulting in a reinforced monsoon-like seasonality (Fig. 9 j,k,l). By preventing the advection of westerly mid-latitudes winds to lower latitudes, it channels this humidity towards inland Asia in summer, thus reducing the aridity



1167 of the region. On the other hand, and although being sometimes put forward as a plausible  
 1168 event contributing to modification of monsoon strength during the Miocene (Bialik et al.,  
 1169 2019, 2020; Rögl, 1997; J. Sun et al., 2021), the mechanical effect of the Tethyan seaway  
 1170 closure, without strong modification of land extension in the Arabian Peninsula region, re-  
 1171 mains limited in our experiments (+1 m/s for the WYI between experiments EMio\_REF  
 1172 and EMio\_TetSw in Fig. 5 d).

## 1173 **6.2 The Eastern Asian monsoon under the crossed influence of Asian orog-** 1174 **raphy, continentality and $p\text{CO}_2$**

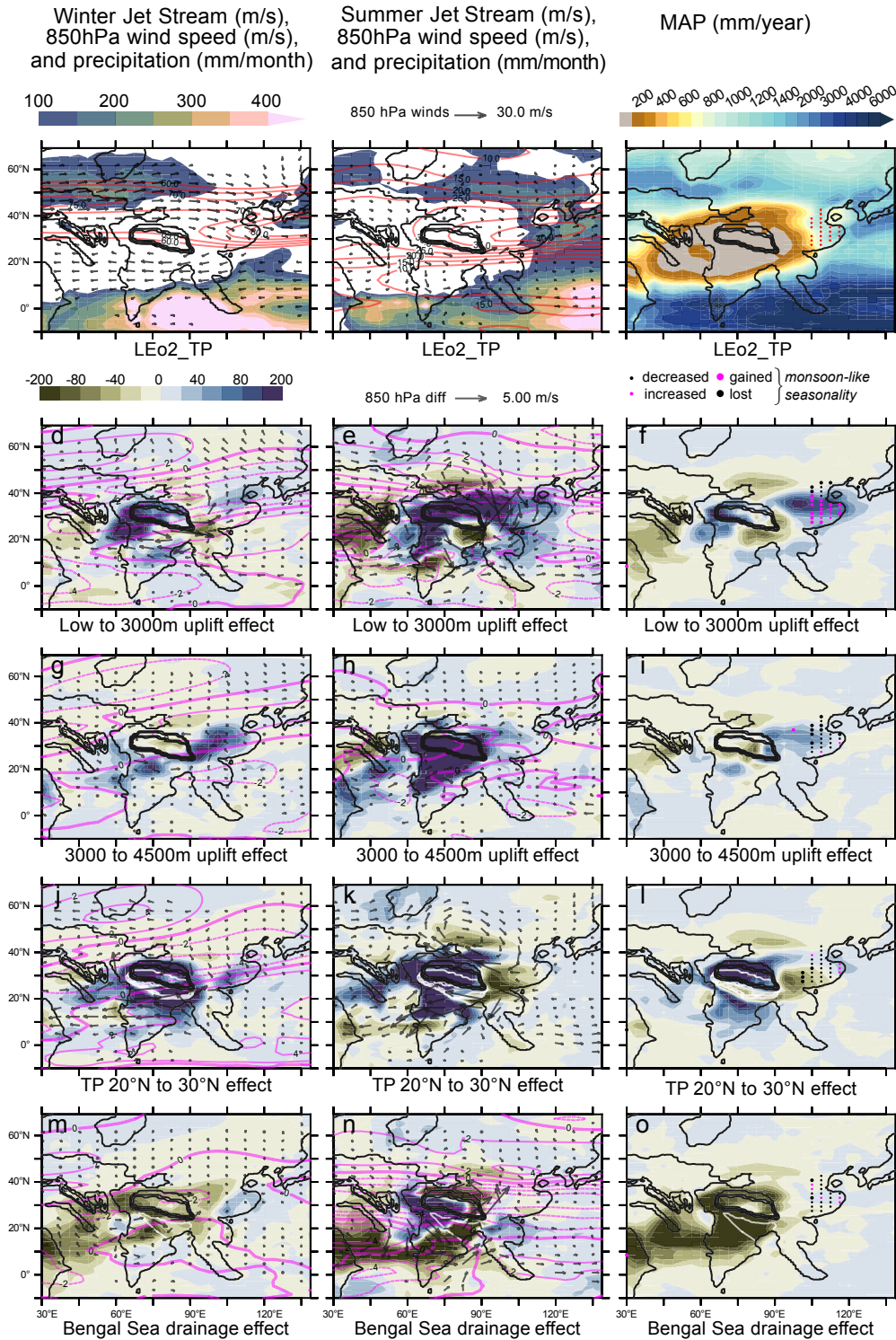
### 1175 ***6.2.1 EASM: precipitation patterns and summer Jet Stream position evo-*** 1176 ***lution***

1177 As opposed to the South Asian region, monsoon-like precipitation seasonality is simu-  
 1178 lated in Eastern Asia in all reference experiments, although it remains initially confined to  
 1179 Northeastern China in the Eocene (Fig. 4 a). As this region of monsoon-like precipitation  
 1180 regime expands to southeastern and inland China throughout the Cenozoic and reaches  
 1181 modern-like extension in the mid Miocene experiment, the Paleogene broad zonal arid band  
 1182 retreats from China (Fig. 4 m). Today, precipitation in eastern China are triggered by the  
 1183 penetration of the Meiyu-Bayu front as it follows the migration of the Jet Stream north of  
 1184 the Tibetan Plateau during summer (Kong et al., 2017). In our simulations, the summer Jet  
 1185 Stream (approximated through the location of maximum zonal wind velocity in the tropo-  
 1186 sphere in August), transits above the proto-Tibetan Plateau and Eastern China in a strong  
 1187 zonal flux in the Eocene and Oligocene experiments (up to 30-35 m/s centered at  $\sim 30^\circ\text{N}$   
 1188 Fig. 4 c,g). It then weakens (10-20 m/s) and progressively migrates north of the Plateau in  
 1189 the Miocene experiments (Fig. 4 k,o,s). The Jet also loses its very zonal configuration and  
 1190 undulates towards Japan. This relocation of the Jet Stream is accompanied by increased  
 1191 precipitation in eastern China, southeastern Siberia and Japan, reaching 200-300 mm in  
 1192 August (Fig. 4 k,o,s), and locally accounting for up to 25% of the annual precipitation in  
 1193 this sole month. These results suggest that the Jet Stream seasonal displacement indeed  
 1194 amplifies the EASM rainy season, although seasonal precipitation have been observed in  
 1195 this region in Late Eocene experiments.

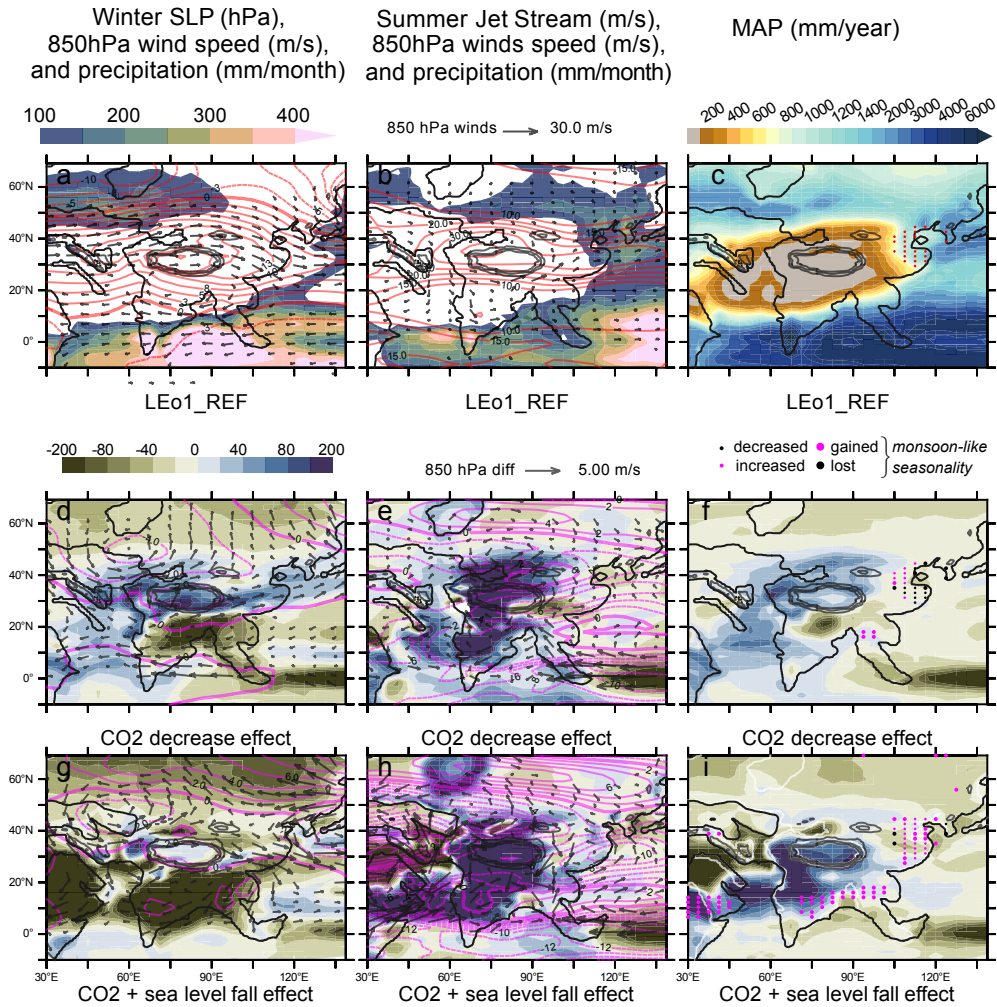
### 1196 ***6.2.2 Summer Jet Stream migration and associated EASM precipitation*** 1197 ***driven by $p\text{CO}_2$ decrease at the EOT and the Northern TP, Tian*** 1198 ***Shan, Pamir and Mongolian uplift in the Miocene***

1199 To identify the drivers responsible for the Jet Stream seasonal migration about the  
 1200 Tibetan Plateau, we here again take advantage of the available sensitivity experiments. All  
 1201 1120 ppm (or 4X) late Eocene simulations display a behavior of the Jet opposite to its  
 1202 modern counterpart, with a northward migration in winter, and a southward migration in  
 1203 summer, below or over the proto-Central TP (Fig. 10 a). This tendency remains relatively  
 1204 undisturbed with alternative TP configurations (Fig. 10 a and Fig. 6 e,h,k,n). The  $p\text{CO}_2$   
 1205 halving in the Eocene-Oligocene reverses this tendency (LEo1\_2X and EOLI\_REF, compared  
 1206 to LEo1\_REF). It induces a migration northward of the Tibetan landform in late summer  
 1207 to early fall (Fig. 10 a and Fig. 7 e,h), and an important decrease in winter zonal speed  
 1208 ( $\sim 50\text{m/s}$  against  $\sim 70\text{ m/s}$  in other Eocene simulations, Fig. 10 a and Fig. SI 21 a to h).  
 1209 This migration is however restricted to the TP region, and the Jet Stream remains at low  
 1210 latitude over eastern Asia ( $\sim 35^\circ\text{N}$ , Fig. 10 b and Fig. SI 19), with few summer precipitation  
 1211 (100-150 mm/month).

1212 All subsequent Miocene simulations display an important latitudinal migration of the  
 1213 Jet along the year, both in the TP region (Fig. 10 c,e), and over eastern Asia (Fig. 10  
 1214 d,f). The northward migration of the Jet is comparable between early Oligocene and early  
 1215 Miocene experiments (up to  $\sim 38\text{-}40^\circ\text{N}$  in September), despite a higher Central Tibet (from  
 1216 3000 to 4000 m) and a wider Tibetan landform, due to the incipient uplift of Northeastern

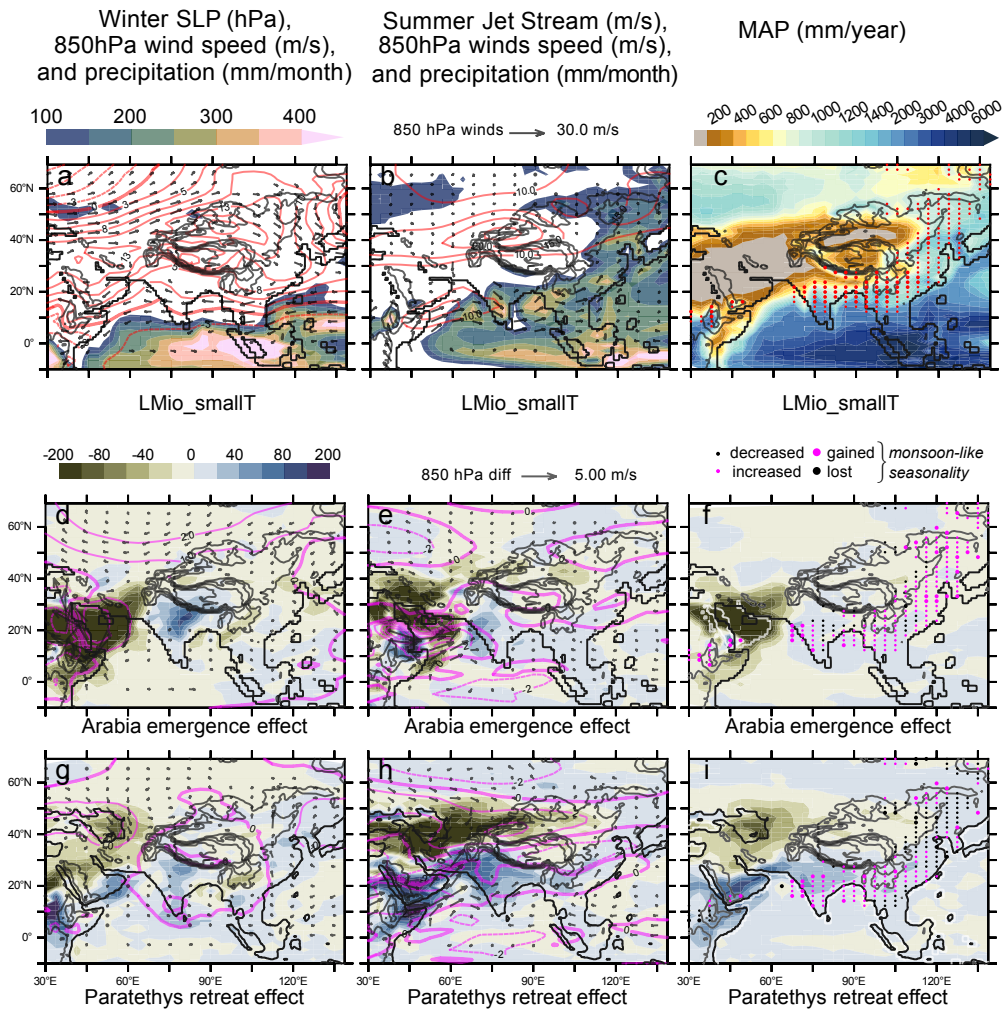


**Figure 6.** Sensitivity to the Tibetan configurations on Eocene climate: (a,d,g,j,m) winter precipitation (shading), 850 hPa winds (vectors), jet stream speed (contour) ; (b,e,h,k,n) on summer precipitation (shading), 850 hPa winds (vectors), jet stream speed (contour); (c,f,i,l,o) mean annual precipitation (shading) and monsoon-like seasonality according to the MPI index (red dots). Sensitivity tests are compared to the simulation LEO2\_TP (a,b,c) and anomalies (d to o) are expressed as "test minus LEO2\_TP", except for the "lowered TP effect" which is "LEO2\_TP minus 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 changes 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 test topography.

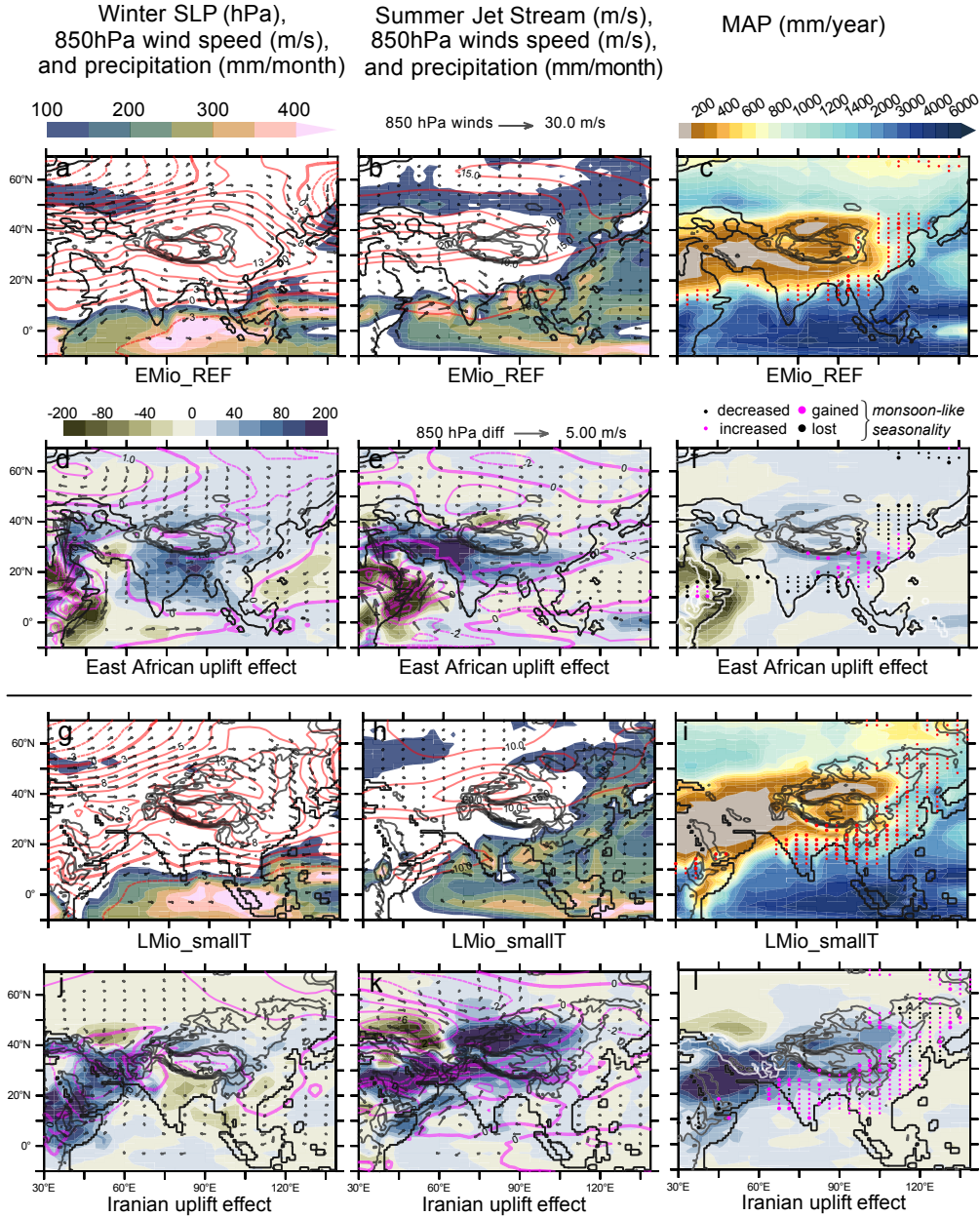


**Figure 7.** Influence of the  $p\text{CO}_2$  decrease and sea level drop at the EOT on: (a,d,g) winter precipitation (shading), 850 hPa winds (vectors), Sea Level Pressure (contour) ; (b,e,h) on summer precipitation (shading), 850 hPa wind (vectors), jet stream speed (contour); (c,f,i) mean annual precipitation (shading) and monsoon-like seasonality according to the MPI index (red dots). Sensitivity tests are compared to the simulation LEO1\_REF (a,b,c) and anomalies (d to i) are expressed in order to highlight the impact of the test (i.e. test-LEO1\_REF). 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): four types of seasonality changes induced by uplift are displayed (e.g. regions switching from no monsoon to monsoon-like seasonality after the sea level fall and  $p\text{CO}_2$  decrease in thick magenta dots) ; white topographic contours indicate the LEO1\_REF contour.





**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 simulation LMio\_smallT (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). 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): four types of seasonality changes induced by continentality are displayed (e.g. regions switching from no monsoon to monsoon-like seasonality after continentality increase in thick magenta dots) ; white topographic contours indicate the test simulation.



**Figure 9.** Influence of the East African uplift in the Early Miocene (a to f) and of Iran-Zagros 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 (Iran-Zagros) uplift is compared to the simulation EMio\_REF (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). 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): four types of seasonality changes induced by continentality are displayed (e.g. regions switching from no monsoon to monsoon-like seasonality after continentality increase in thick magenta dots) ; white topographic contours indicate the uplifted area.

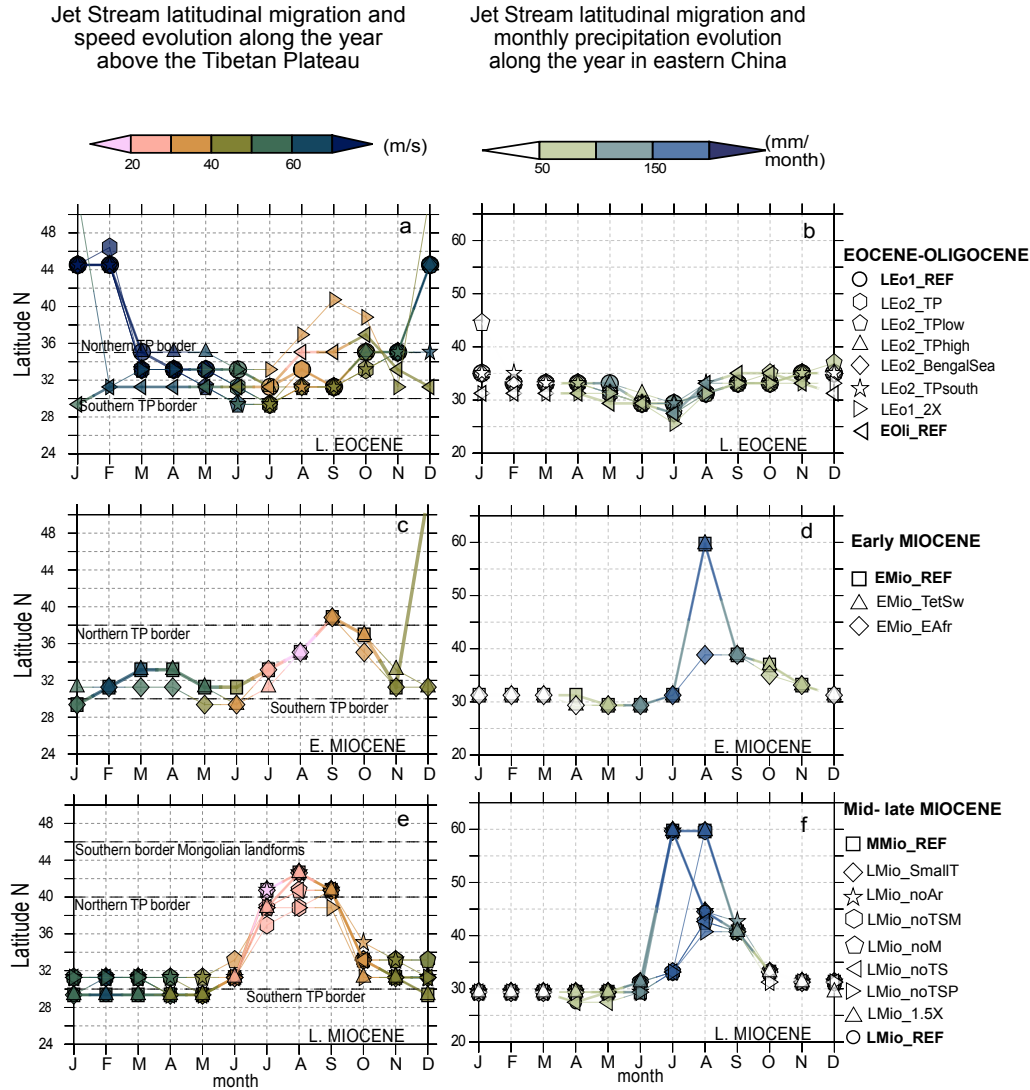
1217 and Northwestern TP portions in the early Miocene simulation (the Tian Shan is  $\sim 1000$  m  
 1218 and the NE Tibet  $\sim 2000$  m high). We therefore suggest that the Jet Stream latitudinal mi-  
 1219 gration is primarily triggered by  $p\text{CO}_2$  drawdown, and resulting tropospheric temperature  
 1220 gradients, and that a significant threshold is crossed at the Greenhouse-Icehouse transi-  
 1221 tion (EOT,  $\sim 34$  Ma). In contrast, the subsequent variations in the Jet Stream latitudinal  
 1222 migration in the course of the Miocene, appear to be mainly driven by topography changes.

1223 The opening of the Tethyan seaway in the early Miocene (comparing EMio\_REF and  
 1224 EMio\_TetSw), the Iran-Zagros landform uplift (comparing LMio\_SmallT and LMio\_REF),  
 1225 or a  $p\text{CO}_2$  decrease (comparing LMio\_REF and LMio\_1.5X) in the late Miocene do not  
 1226 appear to trigger important variations in the Jet Stream migration (Fig. 10 c to f). Other  
 1227 experiments, on the other hand, induce notable variations in response to localized uplift  
 1228 of specific landforms, especially in the mid-late Miocene. Sensitivity experiments removing  
 1229 either Mongolia, Tian Shan, or a combination of both, or Tian Shan and Pamir show that  
 1230 these landforms all contribute to amplify the northward migration of the Jet in summer,  
 1231 both in the TP region and in Eastern Asia (from  $39\text{--}41^\circ\text{N}$  in LMio\_noTS, LMio\_noTSP,  
 1232 LMio\_noTSM and LMio\_noM to  $43^\circ\text{N}$  in MMio\_REF and LMio\_REF, Fig. 10 e and Fig.  
 1233 11 h,k,n). This is especially marked in the experiment removing specifically the Tian Shan  
 1234 orogen, which results in the Jet staying above the TP, without passing its northern edge (Fig.  
 1235 11 h,k,n). We therefore propose that the final uplifts of the Tian Shan, and in a lesser way  
 1236 of the Pamir (from  $\sim 1000$  m in the early Miocene to  $\sim 3000$  m in the mid and late Miocene  
 1237 experiments) and Mongolian mountains (from  $\sim 500$  m in the early Miocene to  $\sim 1500$  m in  
 1238 the mid and late Miocene experiments) are critical to deviate the Jet northward in summer  
 1239 (Fig. 11 e,h,k,n and Fig. SI 19 o to r). This relocation of the Jet Stream in summer induces  
 1240 orographic precipitation on the uplifted landforms, aridification of the regions standing to  
 1241 the west ( $-20$  to  $-60\%$  of summer precipitation in eastern Kazakhstan and Uzbekistan) and  
 1242 increased precipitation in Eastern Mongolia ( $+20$  to  $60\%$ ). This translates into increased  
 1243 monsoon-like seasonality in Northeastern Asia (magenta dots in Fig. 11 f,i,l,o).

### 1244 **6.2.3 East Asian winter monsoon: the Siberian High genesis**

1245 Modern EAWM is characterized by cold and dry climate, together with northwesterly  
 1246 low tropospheric winds above  $20^\circ\text{N}$  and northeasterly winds below  $20^\circ\text{N}$ . This peculiar atmo-  
 1247 spheric circulation is driven by the strong pressure gradients settling between the Siberian  
 1248 High anticyclone, the Aleutian Low and the Maritime Continent Low (L. Wang & Chen,  
 1249 2014; T. Ma & Chen, 2021). Our results suggest strong changes in surface temperature  
 1250 gradients throughout the experiments, that induce an important remodeling of associated  
 1251 pressure and wind patterns (Fig. 12 and Fig. 13). In our Eocene experiment, a high-pressure  
 1252 band centered around  $30^\circ\text{N}$  is simulated, but it remains mainly zonal, following large-scale  
 1253 Hadley winter descending branch driven by latitudinal surface temperature gradients (Fig.  
 1254 12 b). Indeed, the temperature difference between Tibetan region ( $\sim 10^\circ\text{C}$ ) (located below  
 1255 the high pressure cell) and neighboring North Pacific Ocean ( $\sim 30^\circ\text{C}$ ) remains relatively low  
 1256 ( $\sim 20^\circ\text{C}$ , Fig. 12 a), when compared to modern temperature difference at the same season  
 1257 ( $\sim 40^\circ\text{C}$ , Fig. 14). As a result, no significant pressure difference is simulated between the  
 1258 Asian continent and the North Pacific in the Eocene and the calculated EAWM index is  
 1259 low ( $< 5$ , Fig. 13), reflecting only the North-South pressure gradient between Asia and the  
 1260 Maritime Continent.

1261 The extent of the Asian high pressure zone increases gradually in the early Oligocene  
 1262 and early Miocene experiments (Fig. 12 d,f), but its modern-like Siberian High shape is only  
 1263 simulated in the mid and late Miocene experiments, when its center relocates over Siberia  
 1264 ( $\sim 40\text{--}60^\circ\text{N}$ , Fig. 12 h,j). Additionally, the North Pacific low relocates  $10^\circ$  to the south  
 1265 from the early Oligocene simulation onward and steadily decreases until the late Miocene  
 1266 (from over  $\sim 14$  to  $10$  hPa anomaly). All these changes conspire to increase progressively  
 1267 both North-South and East-West pressure gradients, and therefore the global EAWM in-  
 1268 dex, which reaches maximal values in the mid and late Miocene experiments ( $\sim 15$ , Fig.

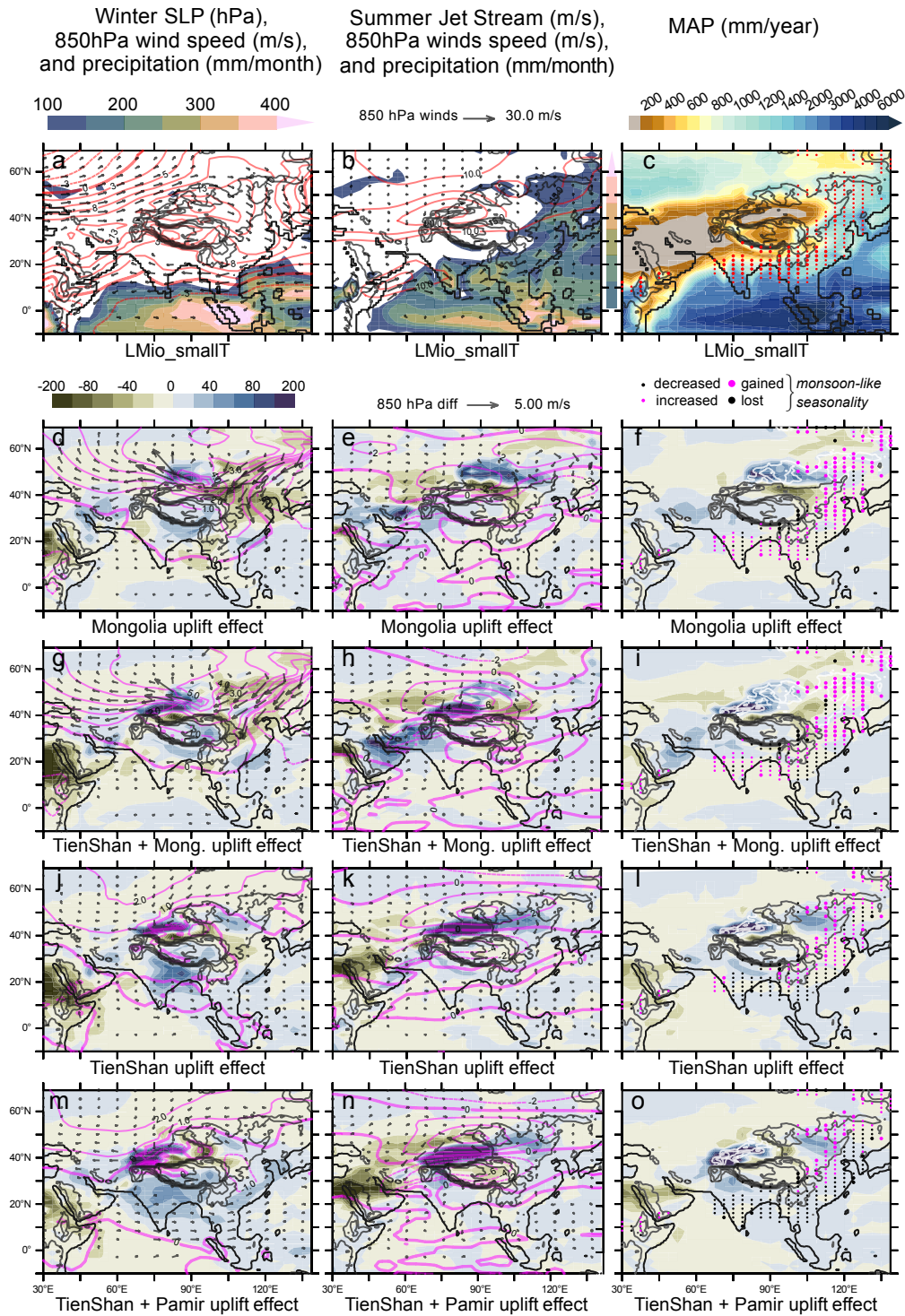


**Figure 10.** Jet Stream behavior along the year over the Tibetan Plateau and eastern China for all sensitivity experiments, divided into (**up**) Late Eocene, (**mid**) early Miocene and (**bottom**) mid to late Miocene. (**left**) Jet Stream latitudinal migration and speed (shading) evolution along 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 (shading) amounts along the year in East Asia (averaged over 100:120°E).

1269 13). Following this reorganization in pressure patterns simulated during the Cenozoic, low  
 1270 tropospheric winds transiting over China display a progressive change in orientation. While  
 1271 their zonal component remains important in the Paleogene and early Miocene experiments  
 1272 (Fig. 12 b,d), the development of the Siberian High triggers a marked increase in their  
 1273 meridional component in the mid and late Miocene experiments (Fig. 12 h,j).

1274 This profound atmospheric circulation reorganization is driven by temperature gradi-  
 1275 ent changes, and mainly due to progressive cooling of the Tibetan, Mongolian and Siberian  
 1276 regions, as illustrated by the  $\sim 22^\circ$  shift southward of the  $0^\circ\text{C}$  isotherm between the late  
 1277 Eocene and the late Miocene simulations (Fig. 12 a,i). Several factors may be responsible





**Figure 11.** Influence of Mongolia, Tian Shan and Pamir regional uplifts on (a,d,g,j,m) winter precipitation (shading), 850 hPa winds (vectors), Sea Level Pressure (contour) ; (b,e,h,k,n) on summer precipitation (shading), 850 hPa wind (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 simulation LMio\_smallT (a,b,c) and anomalies (d to o) are expressed in order to highlight the impact of the uplift (i.e. LMio\_smallT-test). 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 changes induced by uplift are displayed (e.g. regions switching from no monsoon to monsoon-like seasonality after uplift in thick magenta dots) ; white topographic contours indicate the landforms that were added.

1278 for this important winter cooling, such as the  $p\text{CO}_2$  drawdown, local topography changes  
1279 or increased continentality.

#### 1280 **6.2.4 Siberian High inception driven mainly by Mongolia uplift in the** 1281 **Miocene and its impact on precipitation seasonality**

1282 The Eocene sensitivity experiments display a stable zonal high pressure pattern at mid  
1283 latitudes (Fig. SI 18 a to g), with weak response to changing Tibetan orography or  $p\text{CO}_2$   
1284 values (Fig. 13). Likewise, our Miocene sensitivity experiments allow to highlight that the  
1285 winter pressure patterns in Asia (comprising the Siberian High, the North Pacific Low and  
1286 the Maritime Continent Low) are not impacted by African or Iran-Zagros landforms uplifts,  
1287 the Arabian platform configuration, the Paratethys extent, or a  $p\text{CO}_2$  drop from 560 to 420  
1288 ppm (Fig. 13 and Fig. SI 18 i to n and s,t).

1289 On the other hand, sensitivity experiments reveal the impact of the Mongolian land-  
1290 form uplift in the mid-late Miocene (either Mongolia alone, or both Tian Shan and Mongo-  
1291 lia). Indeed, their settlement induces a strengthening of both the Siberian High (up to  $\sim 6$   
1292 hPa, Fig. 11 d,g) and the North Pacific low, resulting in a reinforcement of the East-West  
1293 and North-South pressure gradients, and ultimately of the EAWM index (from  $\sim 8$  to  $\sim 15$ ,  
1294 Fig. 13). Mongolian (and/or Tian Shan and/or Pamir) uplift also results in increased pre-  
1295 cipitation localized over the landforms (Fig. 11 d to o), despite the overall dryness of inland  
1296 Asia region (SI Fig. 16). As Mongolian uplift deviates the winter westerlies flux northward,  
1297 Eastern China is swept by colder and drier winds coming from Siberia, which translates into  
1298 a marked decrease in winter precipitation (up to 80% decrease, Fig. 11 d). The uplift of  
1299 Tian Shan landforms, alone or together with the Pamir, has a lower impact on the global  
1300 winter pressure patterns (Fig. 13), but nonetheless provokes an almost complete drying on  
1301 its eastern flanks (modern Gobi and Taklimakan deserts), as it acts as an orographic barrier  
1302 to westerly moisture flux (Fig. 11 g,j,m).

1303 These changes on winter wind circulation and precipitation amounts, combined to  
1304 those observed in summer and described above, produce contrasted responses in terms of  
1305 monsoon-like precipitation seasonality. As the Mongolian uplift generates a mixed response  
1306 on summer precipitation in eastern China (Fig. 11 e,h), but dries consistently the region in  
1307 winter (Fig. 11 d,g), it induces important spreading of new regions displaying monsoon-like  
1308 precipitation seasonality in eastern China (thick magenta dots in Fig. 11 f,i). At the oppo-  
1309 site, Tian Shan and Pamir uplift, let alone the important year-round increase in orographic  
1310 precipitation, triggers overall increased winter precipitation amounts in Southeastern Asia  
1311 (Fig. 11 j,m), which leads to widespread reduction of precipitation seasonality in the SAM  
1312 domain (Fig. 11 l,o).

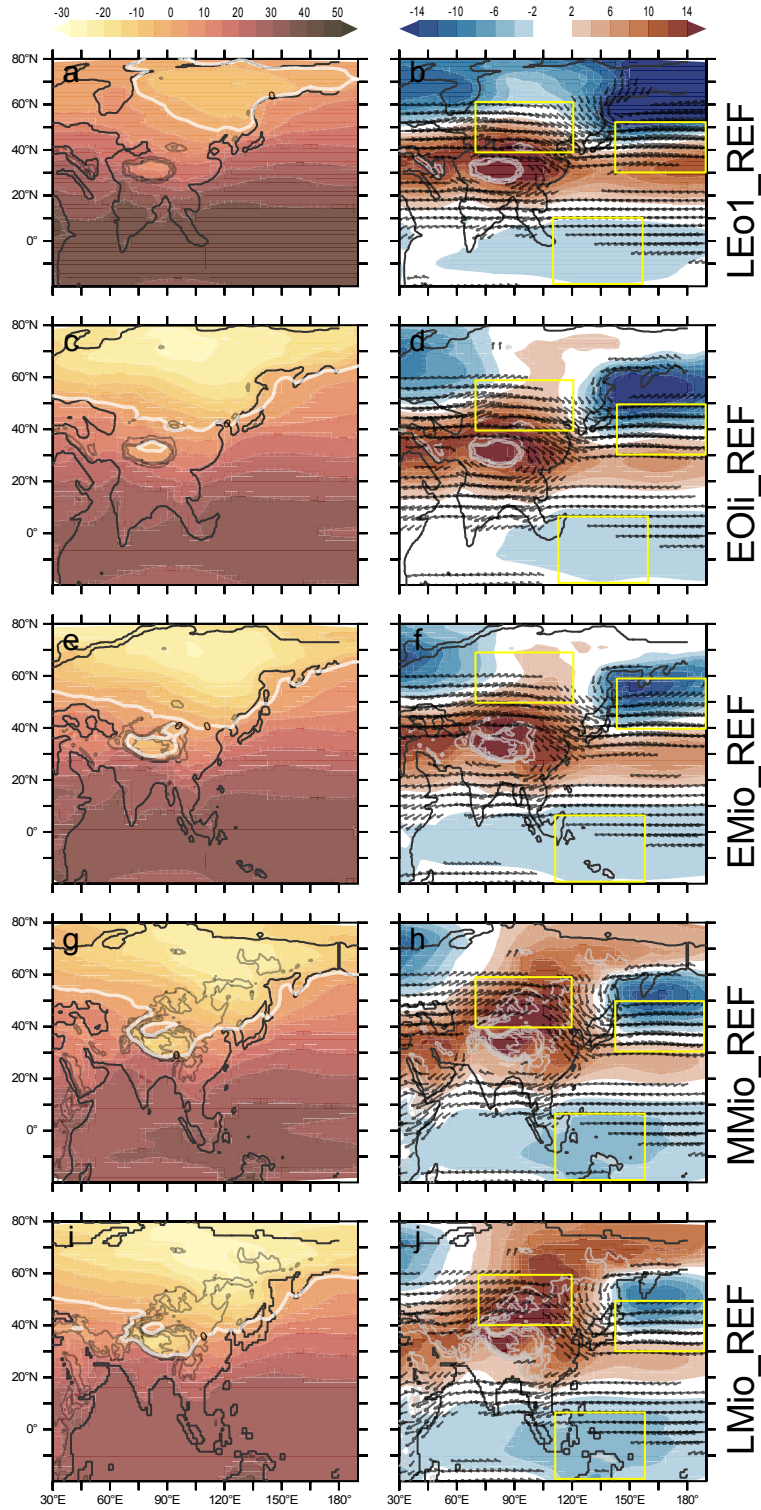
## 1313 **7 Discussion**

1314 A long-standing debate opposes the effect of continentality against orography regarding  
1315 the Asian climate evolution (Prell & Kutzbach, 1992; Kutzbach et al., 1993; Ramstein et  
1316 al., 1997; Fluteau et al., 1999; Z. Zhang et al., 2007; Roe et al., 2016; Zoura et al., 2019;  
1317 R. Zhang et al., 2021). Our results hint that these two features most likely impacted different  
1318 climatic characteristics of the Asian Monsoons and all concurred to the settlement of current  
1319 Asian climate (see Table 6 for a sum-up of the main effects of each forcing on monsoonal  
1320 features). In the following sections we put our main findings in perspective with available  
1321 paleoclimatic indicators and previous modeling work.

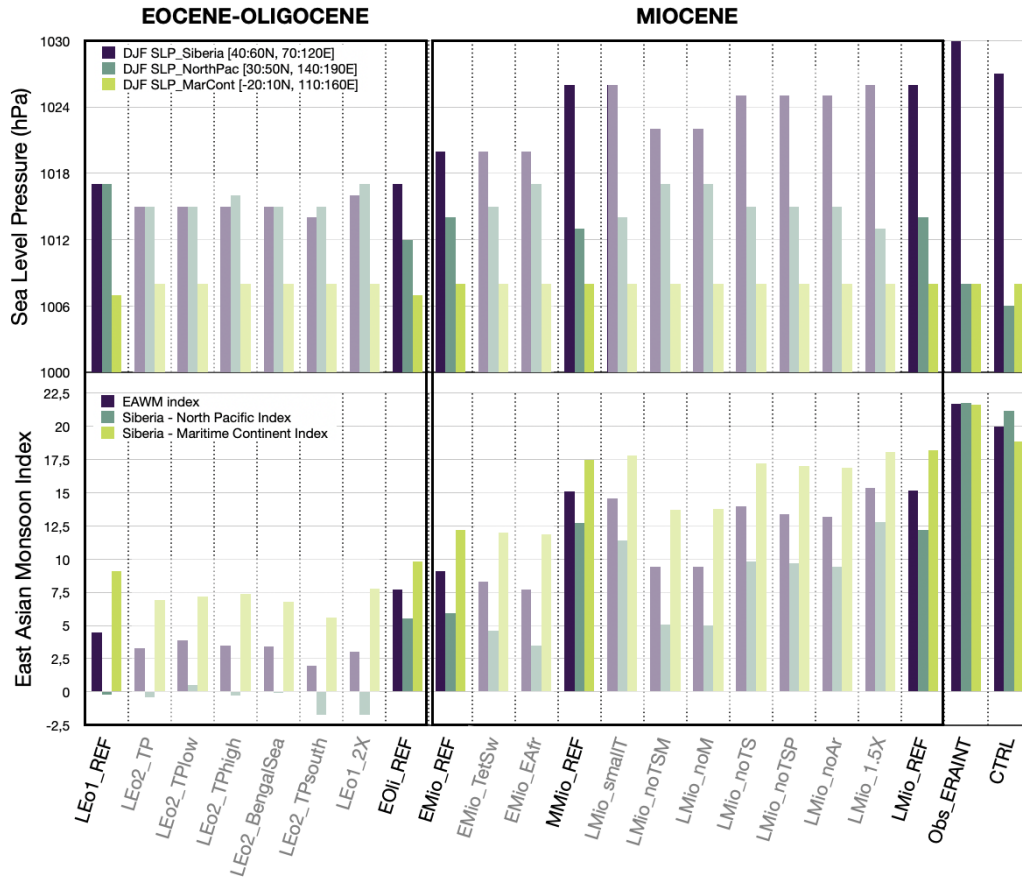
### 1322 **7.1 A possible inception of monsoons in the Paleogene: which dependence** 1323 **to $p\text{CO}_2$ , Tibetan paleogeography and continentality ?**

1324 Our late Eocene set of simulations follows a previous study, in which we introduced  
1325 the results from LEo1\_REF, displaying no monsoon circulation, and discussed a possible dry





**Figure 12.** Evolution of the winter monsoon diagnostics throughout the Cenozoic. **Left column:** December-February (DJF) mean 2m temperature (shading, Celsius), overlain with a white contour line marking the 0°C isotherm ; **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 are calculated the sea level pressure gradients from the East Asian Winter Monsoon Index, presented in section 5.3



**Figure 13.** **Upper panel:** Winter (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).

1326 bias in our simulations, when comparing our results to several proxy localities indicating  
1327 significantly wetter environments (Tardif et al., 2020). Indeed, while markers of arid to sub-  
1328 arid climate are widespread for the late Eocene for example in Northeastern TP (J. X. Li  
1329 et al., 2018; Meijer et al., 2019), Tajikistan (Carrapa et al., 2015) or on the Tarim Basin  
1330 shores (Bougeois et al., 2018), our Eocene experiment fails to render the wetter climate  
1331 inferred from fossil paleoflora and paleofauna in Northern India (Saxena & Trivedi, 2009)  
1332 and Tibet (T. Su et al., 2020; Sorrel et al., 2017). Asian overall aridity in this reference  
1333 Eocene simulation was interpreted as resulting from the hot surface temperatures simulated  
1334 in the region, together with strong mid-tropospheric subsidence that inhibits deep convection  
1335 and condensation, especially over Northern India (Tardif et al., 2020). Nevertheless, much  
1336 wetter and closer to monsoon-like summer climate patterns were simulated in both EAM and  
1337 SAM domains for the late Eocene with the same model and paleogeography, but with orbital  
1338 configurations inducing summer solar maxima (Tardif et al., 2021). This potential orbitally-  
1339 paced wet/dry climate dichotomy is, at least qualitatively, in agreement with stratigraphic  
1340 and paleobotanic evidence from Eastern China (C. Huang & Hinnov, 2019; Tong et al.,  
1341 2002). Another hypothesis for this mismatch was an eventual sensitivity of the simulated  
1342 climate to Asian paleogeography which, as introduced before, is still highly debated for this  
1343 period.

### 1344 ***7.1.1 Late Eocene climate sensitivity to alternative paleogeographic config-*** 1345 ***urations: persistent mismatch with paleoclimate indicators in SAM*** 1346 ***domain***

1347 In this study, we explored the Asian climate sensitivity to different late Eocene Tibetan  
1348 configurations, which aimed at reflecting the main theories regarding the paleogeography at  
1349 that time. We also evaluated the effect of decreased  $p\text{CO}_2$  values (1120 to 560 ppm). Our  
1350 results reveal that the arid pattern previously described shows low sensitivity to the Tibetan  
1351 Plateau configuration or to a reduced  $p\text{CO}_2$ , and is only disturbed by the presence of a  
1352 Bengal Sea (LEo2\_BengalSea), which lowers the surface temperatures and slightly enhances  
1353 precipitation in this region (Fig. SI 16 f). This tendency is however opposite to similar  
1354 modeling studies comparing an early and a late Eocene paleogeography, and which suggest  
1355 increased precipitation after the drainage of this Bengal Sea (X. Li et al., 2017, 2018).  
1356 Comparing these different simulations is nevertheless hampered by the significant difference  
1357 between other boundary conditions that were used. Collectively, these results rather points  
1358 to the high sensitivity of late Eocene climate to surface temperature gradients, at least with  
1359 the IPSL-CM5A2 model, which is further emphasized by the changes induced by the sea  
1360 level fall tested in our Oligocene simulation and presented in the next section.

1361 The EAM region displays higher sensitivity to the paleogeography changes that were  
1362 tested. We simulate that an uplift in the Tibetan region (compared to a low  $\sim 800$  m TP  
1363 in LEo2\_TPLow) would result into an onset or increase in EAM monsoon-like precipitation  
1364 seasonality due to increased summer precipitation (up to 50, Fig. 6e, Fig. SI 16 b,c,d),  
1365 in agreement with previous modeling works (R. Zhang et al., 2018, 2017, 2021; Z. Zhang  
1366 et al., 2007; X. Liu & Yin, 2002; X. Li et al., 2017; Farnsworth et al., 2019; Thomson et  
1367 al., 2021). Likewise, the decreased EAM intensity and increased inland Asian aridity due  
1368 to the northern migration of the TP (its southern edge translating from  $\sim 20$  to  $30^\circ\text{N}$ ) is  
1369 coherent with previous study testing the impact of the latitudinal position of the incipient  
1370 TP (R. Zhang et al., 2017).

### 1371 ***7.1.2 Imprint of the Eocene-Oligocene transition on summer monsoon and*** 1372 ***inland Asia aridification***

1373 Paleoclimatic indicators from inland Asia suggest increased precipitation seasonality  
1374 in EAM and SAM domains, wetter conditions localized in the Tarim and western Qaidam  
1375 basins and overall aridification everywhere else in inland Asia (described in section 3.1.3).  
1376 This aridification was attributed either to global cooling at the EOT (J. Sun et al., 2015;

1377 X. Li et al., 2017), to the Paratethys Sea retreat (Barbolini et al., 2020; Z. Zhang et al.,  
 1378 2007), or both (J. Sun & Windley, 2015). Our sensitivity experiment testing the effect  
 1379 of cooling alone (LEo1.2X) seems to better render wetter conditions in the Tarim-Qaidam  
 1380 region and global drying in the Yunnan region (Fig. 7 f). Taking into account the EOT  
 1381 global sea level drop in addition to global cooling (EOli\_REF) on the other hand, better  
 1382 simulates the monsoon-like precipitation seasonality suggested by paleofloras in both EAM  
 1383 (Vornlocher et al., 2021; R. Spicer et al., 2017; J. Ren et al., 2021; J. Huang et al., 2022;  
 1384 Herman et al., 2017; Ling et al., 2021) and SAM domains (R. Spicer et al., 2017; Srivastava  
 1385 et al., 2012), as well as precipitation amounts in Kazakhstan (Averyanova et al., 2021).  
 1386 Although the comparison to existing modeling studies is limited by the many additional  
 1387 parameters that may vary from our experiments (ice sheets,  $p\text{CO}_2$ , etc.), those which  
 1388 explicitly tested the effect of the Paratethys Sea retreat (R. Zhang et al., 2021; Z. Zhang et  
 1389 al., 2007), tend to simulate similar SAM and EAM precipitation seasonality increase. Our  
 1390 Oligocene experiment however still underestimates precipitation over Tibet (Deng et al.,  
 1391 2021; Song et al., 2020) and does not simulate the drying trend indicated by proxies in the  
 1392 Tajik and Junggar basins as well as in the Gobi region after the EOT (J. Sun & Windley,  
 1393 2015; Baldermann et al., 2021; J. Sun et al., 2015, 2020), although simulated Oligocene  
 1394 mean annual precipitation remain very low in these regions. We suggest this mismatch may  
 1395 be due to the shape of the proto-Tibetan orogen or paleoshorelines used in LEo1\_REF and  
 1396 EOli\_REF, as all of these locations stand very close to the boundary between regions that  
 1397 became indeed drier (Fig. 7). To further explore the impact of land-sea distribution change,  
 1398 the Miocene experiments offer additional input.

### 1399 *7.1.3 Continentality increase as a potential driver of Paleogene wintertime* 1400 *climate variability*

1401 The precocious observation of loess deposits in northeastern Tibet at 40 Ma was inter-  
 1402 preted as resulting from the periodical inception (or reinforcement) of the Siberian High  
 1403 in winter due to enhanced continentality after the Paratethys Sea retreat (Meijer et al.,  
 1404 2021). Other loess deposits are reported during the Oligocene in the Chinese Loess Plateau  
 1405 and Mongolia, apparently coeval with periods of cooling and ice-sheets expansion (J. Sun &  
 1406 Windley, 2015; Wasiljeff et al., 2022). These discoveries question the presence of a Siberian  
 1407 High in the Paleogene, of its role in dust transport and accumulation, and of its forcing  
 1408 factors. Indeed, most modeling studies so far focused on the link between regional uplift  
 1409 and enhanced Siberian High and winter monsoon circulation rather than on the influence  
 1410 of continentality or global climate cooling. Our results suggest that both increased conti-  
 1411 nentality and regional uplift have concurred to the progressive inception of modern Siberian  
 1412 High, while  $p\text{CO}_2$  fall had apparently little impact on its intensity (Fig. 13).

1413 Oligocene global sea level fall increases winter sea level pressure by  $\sim 2$  to 4 hPa over  
 1414 Siberia (Fig. 7 g) and induces the southward expansion of the North Pacific Aleutian  
 1415 low, which altogether increase the pressure gradient between those two regions. Similar  
 1416 trend is obtained, although in a lower magnitude, with our simulation testing the effect  
 1417 of the Arabian platform emergence, while the final retreat of the Paratethys Sea tested in  
 1418 Miocene configuration does not trigger changes (Fig. 13). A modern-like Siberian High with  
 1419 typical EAWM wind patterns is however not simulated until our middle and late Miocene  
 1420 simulations, when the current Tian Shan and Mongolian landforms are set to their present  
 1421 elevations. Collectively, these results suggest that continentality increase may have induced  
 1422 changes in wintertime pressure gradients and atmospheric circulation mostly before the  
 1423 uplift of Tian Shan and Mongolian landforms, but would have had a decreasing importance  
 1424 afterwards.

1425 We also acknowledge that the Mongolian and Tian Shan uplifting history was likely  
 1426 more complex than depicted here, with signs of early uplift in the Hangay region as soon  
 1427 as the Oligocene. Performing additional experiments allowing to test the role of an up-  
 1428 lifted Hangay dome in an Oligocene configuration should allow us to evaluate the potential

1429 presence of a Siberian High at that time. This may help interpreting precocious loess-like de-  
 1430 posits observed at that time in the Chinese Loess Plateau and Mongolia (J. Sun & Windley,  
 1431 2015; Wasiljeff et al., 2022). The presence loess-like deposits in the Xining Basin at  $\sim 40$  Ma  
 1432 however, seems too old to be attributed to the development of the modern Siberian High,  
 1433 if its full development is conditional to uplift of the Mongolian and Tian Shan landforms.  
 1434 This discrepancy suggest that additional mechanisms for dust transport over long distances  
 1435 should be explored for this period, although testing this hypothesis would require further  
 1436 investigation, likely with daily to hourly wind speed outputs (Aoki et al., 2005).

## 1437 **7.2 Peripheral landforms critical role in monsoon seasonal feature settle-** 1438 **ment in the Miocene**

### 1439 ***7.2.1 Tian Shan, Pamir and Mongolia: small mountains, strong effect***

1440 The role played by Mongolian landform uplift in the inception of the Siberian High is  
 1441 one of the most salient outcomes of these experiments. Previous modeling works based on  
 1442 modern geography that have tested the effect of Mongolia uplift (Sha et al., 2015; Shi et al.,  
 1443 2015; Yu et al., 2018) reproduce the same tendencies we have presented, namely a strong  
 1444 increase of the winter sea level pressure over Siberia and typical deflation of the westerly flow  
 1445 towards northern latitudes resulting in a pronounced drying of Eastern Asia and EAWM  
 1446 amplification (Fig. 11 d). Other studies simulate increased EASM precipitation and wind  
 1447 circulation, although they incorporate Tian Shan uplift together with Mongolia (Yu et al.,  
 1448 2018; R. Zhang et al., 2012; H. Tang et al., 2013; Zhisheng et al., 2001), which remains  
 1449 coherent with our findings testing the uplift of both features (Fig. 11 h). Paleoclimate  
 1450 simulations using more realistic paleogeographic reconstructions (Eocene or Oligocene) tend  
 1451 to confirm such tendencies, although the effect of Mongolian landforms was not explicitly  
 1452 tested. Generally, we can observe that when a modern-like Mongolia was prescribed, the  
 1453 simulated winds and precipitation were close to modern-like EAWM (Licht et al., 2014;  
 1454 X. Li et al., 2017), while it is not the case with a reduced Mongolia elevation (Huber &  
 1455 Goldner, 2012).

1456 The effect of the Tian Shan ranges uplift on Asian climate is less conspicuous and  
 1457 seems also more sensitive to the degree of uplift of other surrounding landforms (such as  
 1458 the Pamir-Karakoram, the Hindu Kush or Iran-Zagros), likely owing to its lower latitude.  
 1459 While a rigorous comparison between our results and previous studies is hampered by too  
 1460 many diverging boundary conditions, all tend to point that the Tian Shan uplift would  
 1461 have i) increased yearly orographic precipitation on the ranges, ii) decreased it in the Tarim  
 1462 Basin located in their rain shadow and in inland Asia mid latitudes and iii) reinforced the  
 1463 EAM through increased summer precipitation (Fig. 11 m,n,o) (Baldwin & Vecchi, 2016;  
 1464 H. Tang et al., 2013; Yu et al., 2018; R. Zhang et al., 2017). The key role played by the  
 1465 Tian Shan ranges, together with the Pamirs and Mongolian landforms in triggering the Jet  
 1466 Stream northward migration in summer (Fig. 19e,f and Fig. 9e,h,k,n), as suggested by  
 1467 previous modeling works derived from modern geography (Sha et al., 2020; Shi et al., 2015),  
 1468 likely contributes to increase the EAM. Nevertheless, substantial differences are noted with  
 1469 other modeling studies, especially regarding the effect on SAM domain: where we simulate  
 1470 a weakening of SAM seasonality due to increased winter precipitation, opposite trend is  
 1471 simulated in studies where the Iran-Zagros and Hindu Kush are fully uplifted and prevent  
 1472 the westerly flux to bring moisture to India (Baldwin & Vecchi, 2016; H. Tang et al., 2013;  
 1473 Yu et al., 2018; R. Zhang et al., 2017; X. Wang et al., 2020).

### 1474 ***7.2.2 East Africa and Zagros-Anatolia: the Far West uplifts and SAM*** 1475 ***evolution***

1476 Our set of simulation highlight in addition that paleogeographic evolution in the Middle  
 1477 Eastern and East African regions have been important for the Neogene evolution of the  
 1478 SASM as they strongly contributed to the settlement of modern-like atmospheric circulation



1479 as discussed by Sarr et al. (2022). The presence of intermediate to high elevation topography  
 1480 in East Africa in the early Miocene is for example crucial for the "early" bending of the  
 1481 Somali Jet, before the closure of the West Tethys seaway and settlement of Iran-Zagros  
 1482 topography with modern-like elevation (Fig. 9).

1483 Increasing elevation in the East African region is also responsible, in our simulations  
 1484 for increasing SAWM precipitation (Fig. 9), which contrast with previous studies that sim-  
 1485 ulate reduced SASM precipitation instead (H.-H. Wei & Bordoni, 2016; Chakraborty et al.,  
 1486 2009) as topography would block the advection of moisture from equatorial Africa towards  
 1487 India. We hypothesize that such difference may come from these studies using present-day  
 1488 paleogeography, and/or unchanged fixed SSTs for all the experiments, thereby neglecting  
 1489 the potential effect of the East Africa uplift on Indian Ocean SSTs and SST gradients. These  
 1490 studies however critically highlight that a reinforced Somali Jet may not be systematically  
 1491 correlated with increased SAM precipitation, although this genetic relationship is commonly  
 1492 found in the literature. Alongside, increasing land exposure in the Arabian platform region  
 1493 in the middle Miocene, the uplift of the Iran-Zagros topography strongly reinforces the So-  
 1494 mali Jet speed and northward migration, together with the summer precipitation amounts  
 1495 (Fig. 5, Fig. 9), as previously suggested by (R. Zhang et al., 2017; Acosta & Huber, 2020;  
 1496 H. Tang et al., 2013).

1497 Given their respective impact in the SAM, knowledge of the evolution of those western  
 1498 topographies, although seldom taken into account, is critical to interpret paleoenvironmental  
 1499 indicators. This is especially true for inland Asia and northwestern India regions, because  
 1500 of their key location with respect to the westerlies flux. Westerlies may indeed be deviated  
 1501 either to the North or the South depending on the Zagros-Iran and Tian Shan respective  
 1502 elevation. Indication of increased SAM precipitation during the Miocene (X. Yang et al.,  
 1503 2020) could for example be, at least partly due to Zagros-Iran uplift, as geological evidences  
 1504 indicate that the high topography settled at that time (Austermann et al., 2017; François  
 1505 et al., 2014).

### 1506 **7.3 Miocene continentality increase as an amplifier of SAM, EAM and** 1507 **inland aridity**

1508 The northward motion of the ITCZ over the Arabian Sea and the stronger SASM  
 1509 precipitation in response to continentality increase in the Miocene is a feature clearly sim-  
 1510 ulated in our experiments, and which is obtained with either the Paratethys Sea retreat or  
 1511 the Arabian peninsula emergence (Fig. 8 e,h). The role of the Paratethys Sea retreat as  
 1512 a vector of amplified summer low sea level pressure zone and resulting moisture advection  
 1513 via strengthened Somali Jet is well established by modeling experiments (Ramstein et al.,  
 1514 1997; Fluteau et al., 1999; R. Zhang et al., 2021; Z. Zhang et al., 2007). The Paratethys Sea  
 1515 retreat also drastically dries the "westerly corridor" (centered around 40°N, Fig. 8 h) which  
 1516 overall results in increased inland Asian aridity and moderately increases EAM seasonality,  
 1517 north and south of this latitudinal band (Fig. 8 i). These patterns are mostly coherent with  
 1518 those obtained by previous modeling studies (R. Zhang et al., 2021; Z. Zhang et al., 2007),  
 1519 although here again, important differences in the paleogeography and type of model that  
 1520 were used prevent a rigorous comparison.

1521 On the other hand, the effect of increased land exposure in the Arabian peninsula  
 1522 region on Asian climate was, to our knowledge, significantly less explored by modeling  
 1523 studies and is never mentioned as a potential forcing by field work. In most modeling  
 1524 studies, the region usually evolves together with the global paleogeography, without being  
 1525 specifically tested. We putatively attribute this to the lack of constraints and data in this  
 1526 region. Given the shallowness and extension of the Arabian peninsula (~3 million square  
 1527 meters, roughly comparable to India area), it is very likely that the episodes of partial or  
 1528 complete flooding that it witnessed from the Oligocene to the late Miocene (Barrier et al.,  
 1529 2018; Otero & Gayet, 2001) may have had great consequences on atmospheric circulation

1530 patterns. Our results show that increased peninsula exposure reinforces the Somali Jet  
 1531 in summer and enhances summer precipitation in SAM and EAM domains (Fig. 8 e).  
 1532 Additionally, it drives the onset of upwelling in the Arabian Sea (see Sarr et al. (2022),  
 1533 with the same set of simulations), which are traditionally used as proxies indicating the  
 1534 establishment of a monsoon regime in South Asia. These findings imply that a proto-Somali  
 1535 Jet and intense summer precipitation settled before the emergence the Arabian peninsula in  
 1536 the middle Miocene, and therefore before the recording of upwelling in the Arabian Sea. In  
 1537 any cases, the onset of upwellings in the Arabian Sea was likely decorrelated from potential  
 1538 Miocene uplifts events in the Himalaya-Tibet region (Sarr et al., 2022).

#### 1539 **7.4 Persistent uncertainties and directions for further works**

1540 Our study demonstrates the key role played by land-sea distribution and by all to-  
 1541 pographic features surrounding the Indian Ocean in shaping the modern SAM, EAM and  
 1542 inland Asia aridity. Nevertheless, additional work is required from both modeling and the  
 1543 field research to help refine our understanding of the complex mechanisms at play through-  
 1544 out the Cenozoic in Asia.

##### 1545 **7.4.1 *Opposite effect and thresholds crossings: the need for improved rel- 1546 ative and absolute timing and elevation constraints on uplift events***

1547 Overall, our results highlight the strong crossed influence of the various sub-geographical  
 1548 units on resulting monsoons and therefore the conceptual limitations of performing tests  
 1549 based on modern geographies or assuming homogeneous Asian orography uplift in the course  
 1550 of time to evaluate past monsoon evolution. In that aspect, it is not surprising that most  
 1551 of such studies simulate somehow modern-like summer and winter monsoons when keeping  
 1552 uplifted African, Mongolian and Iranian landforms, as well as an emerged Arabia (Roe et  
 1553 al., 2016; Zoura et al., 2019; Acosta & Huber, 2020).

1554 The need to better constrain the evolution in time, elevation and shape of the various  
 1555 Asian, East African and Anatolian landforms, appears paramount, in order to refine our  
 1556 understanding of Cenozoic paleoclimate evolution. Our results also suggest that some pa-  
 1557 leogeographic events may have had contrary influences on either SAM or EAM, although  
 1558 we might picture them as roughly coeval. This is for example the case of the Tien-Shan  
 1559 and Pamir uplifts that weaken SAM seasonality, while Mongolian and Iran-Zagros uplifts,  
 1560 together with continentality strengthen it. Here, we chose to perform our mid-late Miocene  
 1561 sensitivity tests based on a halfway uplifted Iran-Zagros orogen (LMio\_SmallT), but it is  
 1562 likely that we may obtain different results if testing the uplift on Mongolia (for example) in a  
 1563 paleogeography with an already fully uplifted Iran-Zagros topography. Refining the relative  
 1564 chronology of these various paleogeographic events is therefore critical to better interpret  
 1565 the variations in monsoon intensity recorded by the proxies. Additionally, paleoelevation  
 1566 thresholds are likely to be critical for constraining the Asian climate evolution, due to in-  
 1567 teractions of the topography with either low level (e.g. the summer monsoon winds) or mid  
 1568 tropospheric (e.g. the Jet Stream) air masses (X. Liu & Yin, 2002; X. Wang et al., 2020;  
 1569 Zhisheng et al., 2001; Prell & Kutzbach, 1992).

##### 1570 **7.4.2 *Land-sea distribution and gateway closure***

1571 A more precise view of land-sea distribution and regional seaways evolution is also  
 1572 paramount. Indeed, they influence the surface temperature gradients, and therefore con-  
 1573 dition sea level pressure patterns, low tropospheric circulation as well as oceanic surface  
 1574 circulation. Although our results point to a minimal effect of the mechanical closure of the  
 1575 Tethyan Seaway on regional results, other features need to be tested. For example, a better  
 1576 understanding of the Indonesian Gateway and Indonesian throughflow evolution would be  
 1577 of great interest, as they control part of Indo-Pacific Warm Pool SSTs distribution (Cane  
 1578 & Molnar, 2001). While the final step of the gateway constriction likely occurs during

1579 the Pliocene (Auer et al., 2019), its paleobathymetric and geometric configuration remains  
1580 however largely undocumented between 17-3.5 Ma, due to difficulties in reconstructing the  
1581 complex configuration of the region (Kuhnt et al., 2004). Although no major tectonic defor-  
1582 mation is supposed to have occurred during this period, the shallow depth of the gateway  
1583 likely made it sensitive to middle Miocene sea level fluctuations (Sosdian & Lear, 2020),  
1584 with possible impact the Eastern Indian and Western Pacific oceans surface temperatures  
1585 and therefore on SST gradients.

## 1586 8 Conclusions

1587 In this study, we took advantage of the rich discoveries that have allowed to extend  
1588 the fossil record to the Paleogene in Asia, as well as the consolidation of the understanding  
1589 of the physical concepts underpinning the monsoons brought by modeling studies. Overall,  
1590 we simulate an intensification of all considered monsoon metrics throughout the Cenozoic,  
1591 in response to specific forcings. We find that, in agreement with paleoclimate indicators,  
1592 modern-like monsoons, in terms of atmospheric circulation, seasonality and precipitation  
1593 amounts, likely exist since at least the mid-late Miocene. We however show that high  
1594 seasonality in precipitation may have existed since the Paleogene in both SAM and EAM  
1595 domains, before the settlement of modern atmospheric circulation patterns. In agreement  
1596 with previous studies, our results highlight the rather limited effect of  $p\text{CO}_2$  when compared  
1597 to paleogeographic changes, with the notable exception of the effect of the  $p\text{CO}_2$  drawdown  
1598 at the EOT on the Jet Stream latitudinal migration over the year in the Tibetan region.

1599 First, although the exact shape and extent of coastal shallow environments remain  
1600 poorly constrained, our results strongly hint that any variations in the land-sea distribution,  
1601 whether due to tectonic activity, or sea level fluctuations, would have strongly impacted the  
1602 Asian climate. Many episodes of increased continentality are documented in the course of the  
1603 Cenozoic, including the regional regressions of the Paratethys epicontinental Sea, periodical  
1604 exposure of the Arabian platform due to sea level fluctuations, and global sea level fall  
1605 at the Eocene Oligocene Transition. All of these events likely favored the enhancement of  
1606 summer moisture penetration into Asia, by promoting the extension of the wide subtropical  
1607 low pressure belt and thus amplifying the seasonal latitudinal migration of the ITCZ on  
1608 land.

1609 Second, while the HTP complex was recurrently put forward as the leading mech-  
1610 anism in Asian climate evolution, our results rather points toward a shared responsibility  
1611 between all landforms surrounding the modern Indian Ocean. We also stress the fact that all  
1612 paleogeographic features do not bear the same importance on SASM, EASM or EAWM re-  
1613 spectively. The Tibetan Plateau uplift likely favored the onset of monsoon-like precipitation  
1614 seasonality in eastern Asia in the Eocene, by promoting the advection of moist air masses  
1615 in summer, and by partially shielding the region from westerly moisture input in winter.  
1616 The Miocene Eastern African and Iran-Zagros uplift were critical to intensify and redirect  
1617 the Somali Jet towards the Indian subcontinent, and thus develop the modern-like South  
1618 Asian monsoon summer circulation and amplify moisture advection to Asia in summer.  
1619 Additionally, Iranian-Anatolian uplift, by blocking moisture advection from the westerlies  
1620 to lower latitudes likely contributed enhance winter aridity in most of Southeastern Asia.  
1621 The Himalaya-Tibet complex then conditions the SASM precipitation distribution over the  
1622 South Asian region, triggers heavy orographic precipitation and upper atmosphere heating  
1623 over northern India.

1624 We show that the uplift of the Tian Shan and the Mongolian mountains was key for  
1625 the settlement of the winter Siberian high pressure cell, and resulting aridification of inland  
1626 and eastern Asia in winter. The summer Jet Stream migration northward of the Tibetan  
1627 Plateau was likely amplified in the mid-late Miocene, as the Pamir and Tian Shan landforms  
1628 approached their modern elevation and contributed to its channeling toward higher latitudes.  
1629 This relocation of the Jet Stream amplifies the penetration of moist air masses over eastern

1630 China, thus contributing to increase the EAM precipitation seasonality. Overall, our results,  
1631 in agreement with available paleoclimate indicators and most previous modeling studies,  
1632 show the fundamentally polygenetic history of monsoonal evolution, suggesting that the very  
1633 concept of a "monsoon onset" is inappropriate. We also stress the need for more accurate  
1634 constrains regarding paleoshorelines evolutions, relative timing of uplift and landforms paleo-  
1635 elevations in order to propose better global climate evolution scenarios.

## 1636 Acronyms

1637 **ITCZ** Inter Tropical Convergence Zone  
1638 **SASM** South Asian Summer Monsoon  
1639 **EASM** East Asian Summer Monsoon  
1640 **EAWM** East Asian Winter Monsoon  
1641 **HTP** Himalaya-Tibetan Plateau  
1642 **IGLLJ** Indo-Gangetic Low Level Jet  
1643 **MMCO** Mid Miocene Climatic Optimum  
1644 **EOT** Eocene Oligocene Transition

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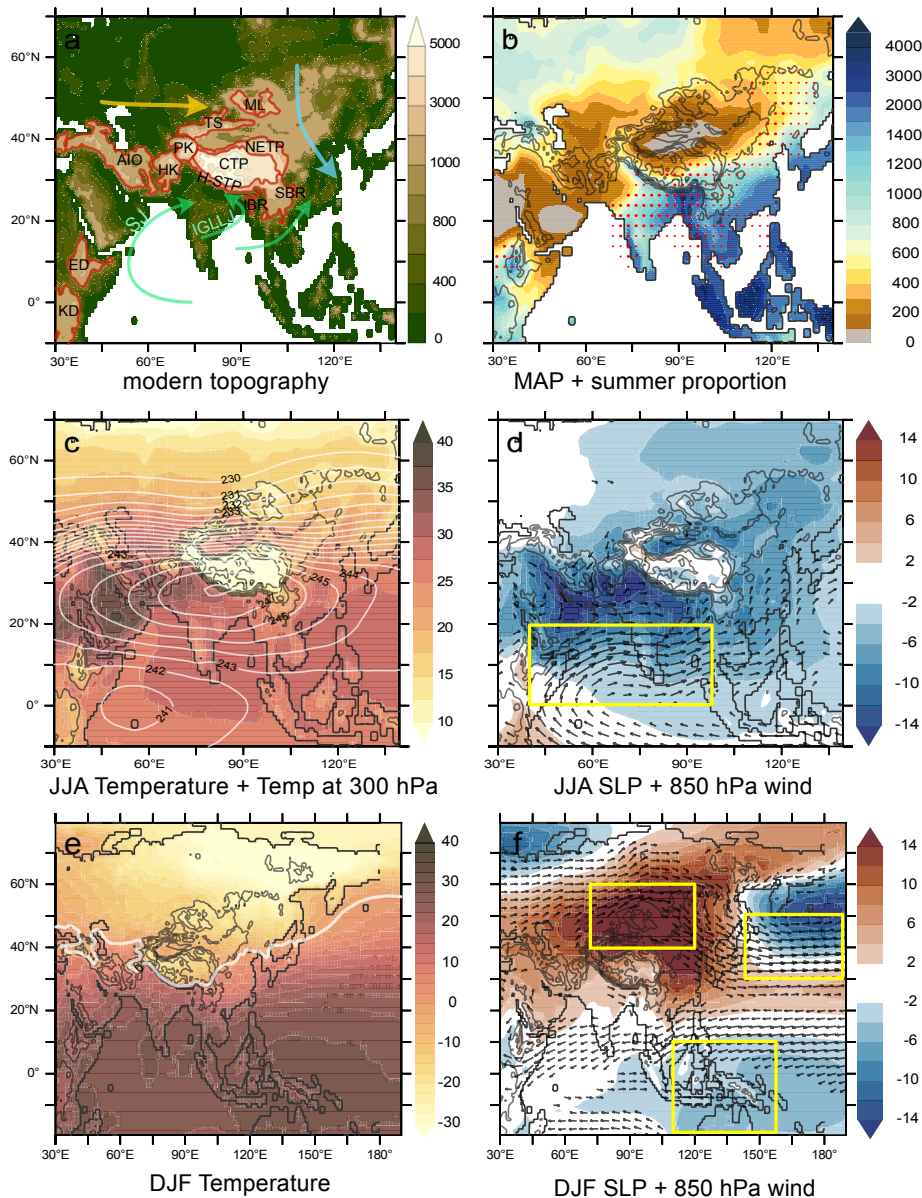
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## 9 Supplementary



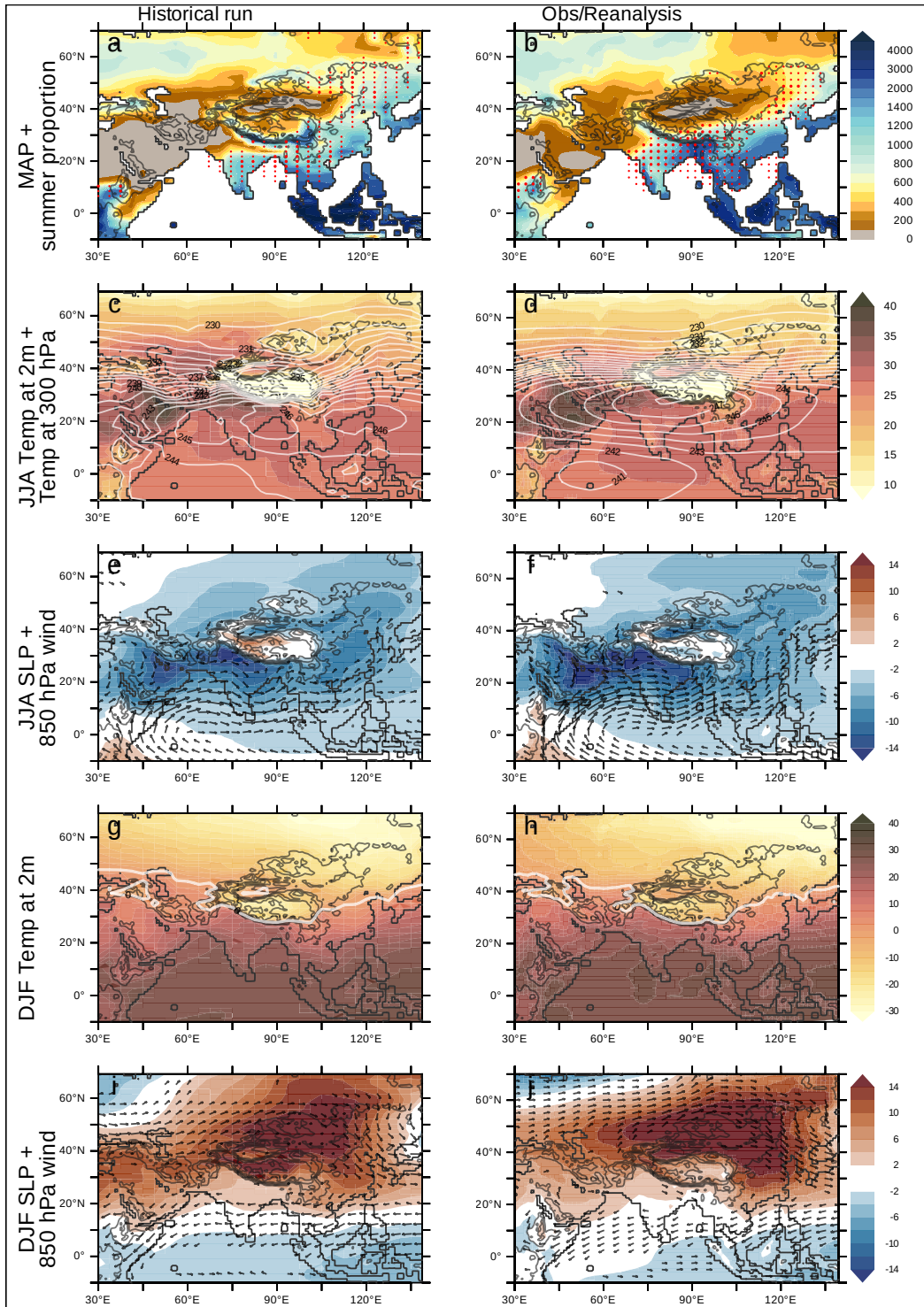
**Figure 14.** South Asian orography and main seasonal characteristics of South and East Asian Monsoons subsystems from observations and reanalysis. (a) Modern geography and topography (in m), with main seasonal winds: Westerlies (yellow arrow), Winter monsoon (blue arrow) and Summer monsoon (green arrows); Somali Jet (SJ), Indo-Gangetic-Low Level Jet (IGLLJ) and main geographic units mentioned in Section 4 highlighted: KD - Kenyan Dome, ED - Ethiopian Dome, AIO - Anatolian and Iranian orogens, PK - Pamir-Karakoram, HK - Hindu Kush, CTP - Central Tibetan Plateau, H-STP - Himalayas and Southern Tibetan Plateau, NETP - North-eastern TP, TS - Tian Shan, IBR - Indo Burman Ranges, SBR - Sino Burman Ranges, ML - Mongolian Landforms.; (b) Mean Annual Precipitation (shading, mm/yr), and regions where the Monsoon Precipitation Index is over 0.5 (thin red dots) and 0.75 (thick red dots). The Monsoon Precipitation Index is defined in the Monsoon Metrics section 5.3; (c) JJA mean 2m temperature (shading, Celsius) and 300 hPa atmospheric temperature (white contour, Kelvin); (d) JJA normalized Sea Level Pressure anomalies (shading, hPa) and 850 hPa winds (vectors over 4m/s). Yellow box highlight the region of calculation of the Webster-Yang Index, defined in the Monsoon Metrics section 5.3; (e) DJF mean 2m temperature (shading, Celsius); (f) DJF normalized Sea Level Pressure anomalies (shading, hPa) and 850 hPa winds (vectors over 4m/s). Yellow boxes highlight the regions of calculation of the Eastern Asian Monsoon Index, defined in the Monsoon Metrics section. In all captions, topography (each 1000 m) is in gray contour. Precipitation for 1979-2021 from the Global Precipitation Climatology Project (GPCP) (Pendergrass, Angeline, Wang, Jian-Jian National Center for Atmospheric Research Staff (Eds). Last modified 06 Nov 2020. "The Climate Data Guide: GPCP (Monthly): Global Precipitation Climatology Project." Retrieved from

	Reference
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8	(J. Sun & Windley, 2015)
9	(Baldermann et al., 2021)
10	(Caves Rugenstein et al., 2014)
11	(Kent-Corson et al., 2009)
12	(Zheng et al., 2015)
13	(Meijer et al., 2021)
14	(Licht et al., 2014)
15	(Wasiljeff et al., 2022)
16	(Hoorn et al., 2012)
17	(X. Ma & Jiang, 2015)
18	(Guo et al., 2002)
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31	(Q. Wang et al., 2021)
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33	(Zheng et al., 2022)
34	(Sorrel et al., 2017)
35	(Jacques, Guo, et al., 2011)
36	(Xia et al., 2009)
37	(Xing et al., 2012)
38	(Wan et al., 2007)
39	(Jia et al., 2003)
40	(P. Clift et al., 2008)
41	(P. D. Clift, 2020)
42	(Z. Ding et al., 2021)
43	(Srivastava et al., 2012)
44	(Bhatia, Khan, et al., 2021)
45	(Shukla et al., 2014)
46	(Khan et al., 2014)
47	(Vögeli et al., 2017)
48	(Bhatia et al., 2022)
49	(Bhatia, Srivastava, et al., 2021)
50	(Srivastava et al., 2018)
51	(Sanyal et al., 2004)
52	(S. Ali et al., 2021)
53	(Betzler et al., 2016)
54	(Betzler et al., 2018)
55	(Beasley et al., 2021)
56	(Zhuang et al., 2017)
57	(Page et al., 2019)
58	(Q. Yuan et al., 2021)
59	(Meng et al., 2018)

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

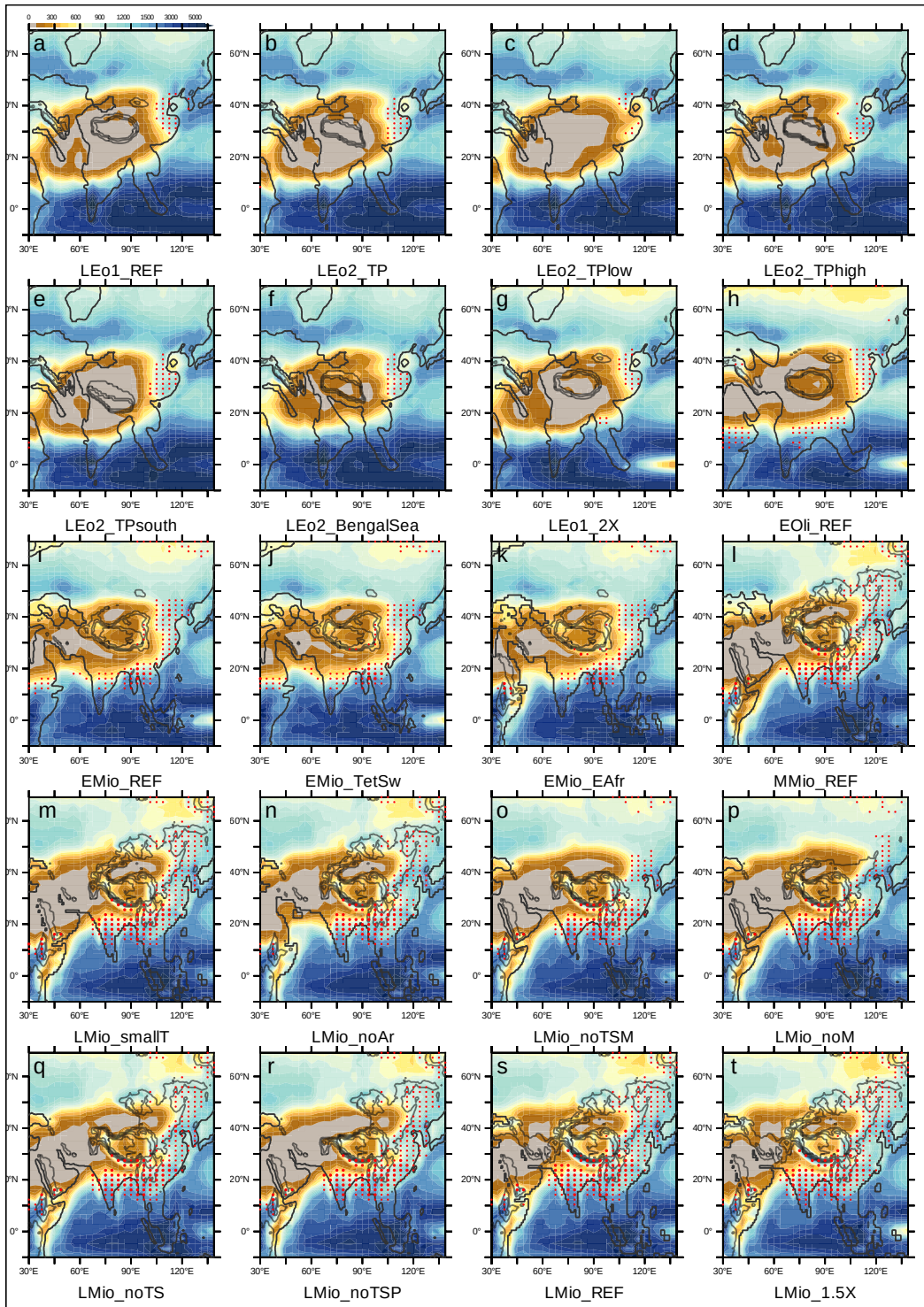
	Reference (suite)
60	(Miao, 2013)
61	(Z. Zhang & Sun, 2011)
62	(J. Sun & Zhang, 2008)
63	(Q.-g. Sun et al., 2002)
64	(J. Sun et al., 2020)
65	(T. Su, Spicer, et al., 2019)
66	(Wu et al., 2021)
67	(Song et al., 2020)
68	(Miao, Wu, et al., 2016)
69	(D. Rea et al., 1985)
70	(Pettke et al., 2002)
71	(Ziegler 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)
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97	(Z. Liu et al., 2014)
98	(Miao et al., 2011)
99	(Charreau et al., 2012)
100	(A. E. 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)
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**Table 3.** (suite) List of references used in Figure 2 for paleoclimate indicators compilation

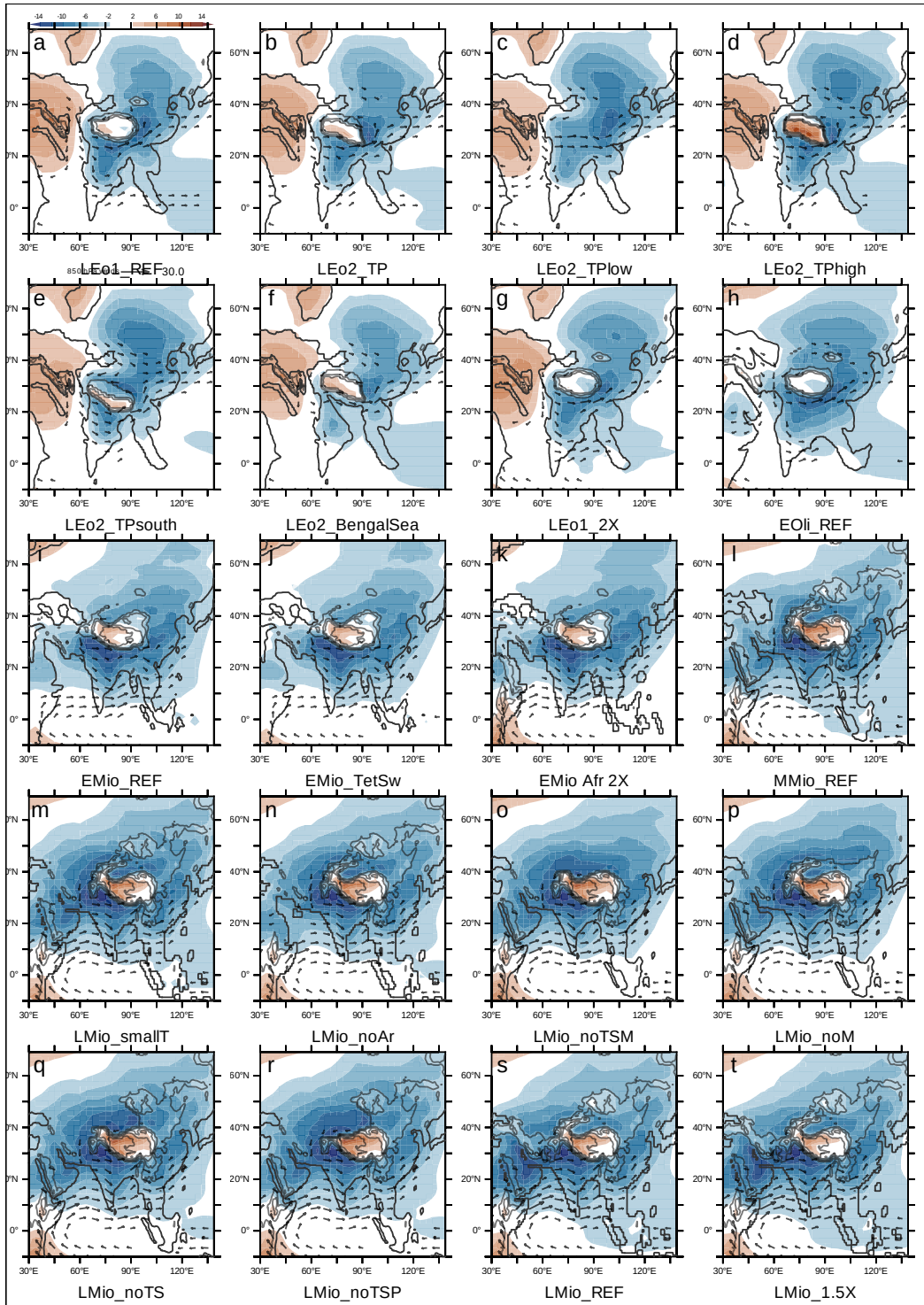


**Figure 15.** Comparison of monsoon diagnostics obtained in our control historical simulation (**Left column**) with modern observations and reanalysis (**Right column**). (**a,b**) Mean Annual Precipitations (shading, mm/yr), and regions where the Monsoon Precipitation Index is over 0.5 (thin red dots) and 0.75 (thick red dots) ; (**c,d**) JJA mean 2m 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 4m/s); (**g,h**) DJF mean 2m temperature (shading, Celsius) ; (**i,j**) DJF normalized Sea Level Pressure anomalies (shading, hPa) and 850 hPa winds (vectors over 4m/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.

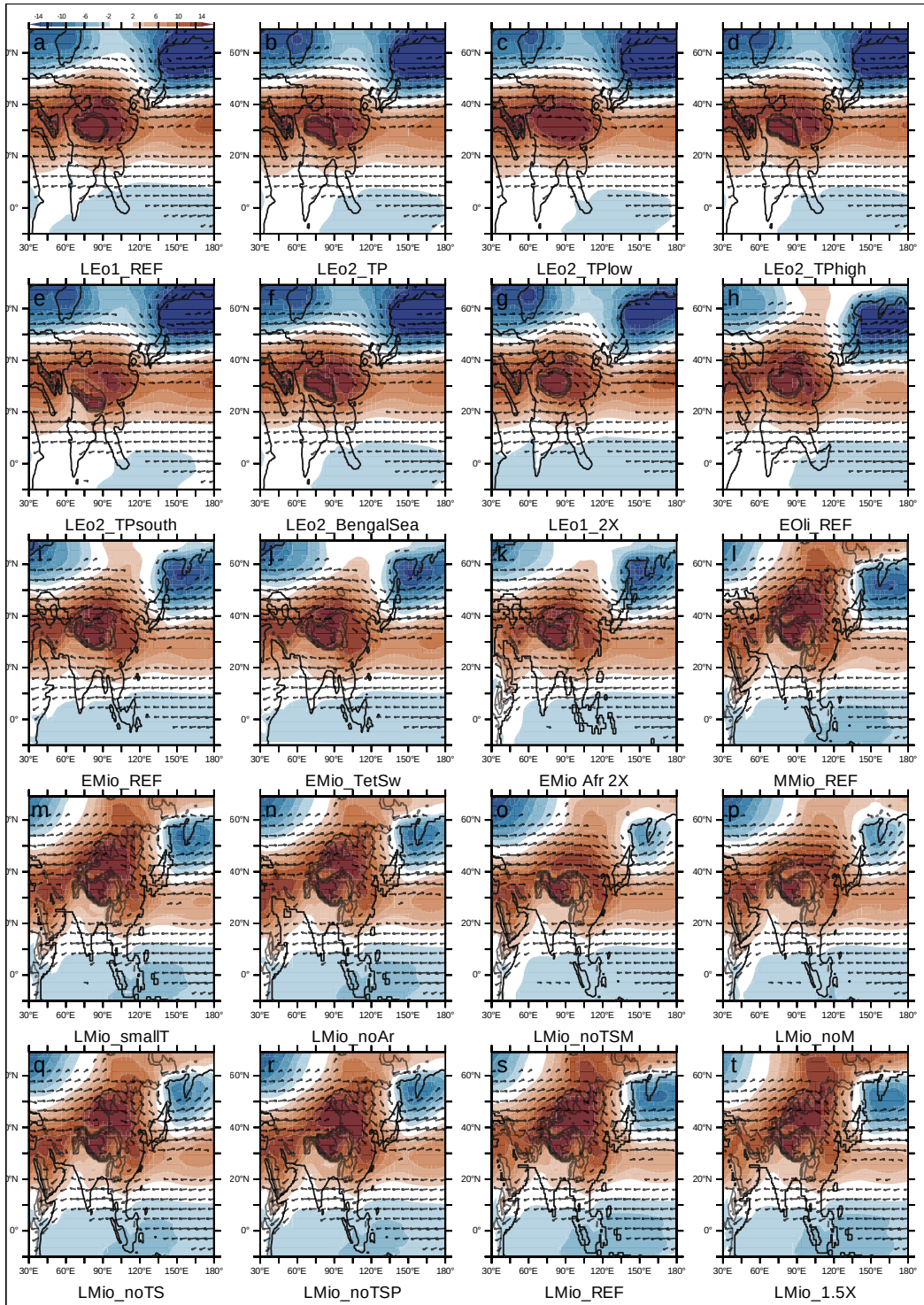




**Figure 16.** Mean Annual Precipitation (shading, mm/yr), overlain with regions where the Monsoon Precipitation Index (B. Wang & Ding, 2008) is over 0.5 (thin red dots) and 0.75 (thick red dots), for all reference and sensitivity experiments.

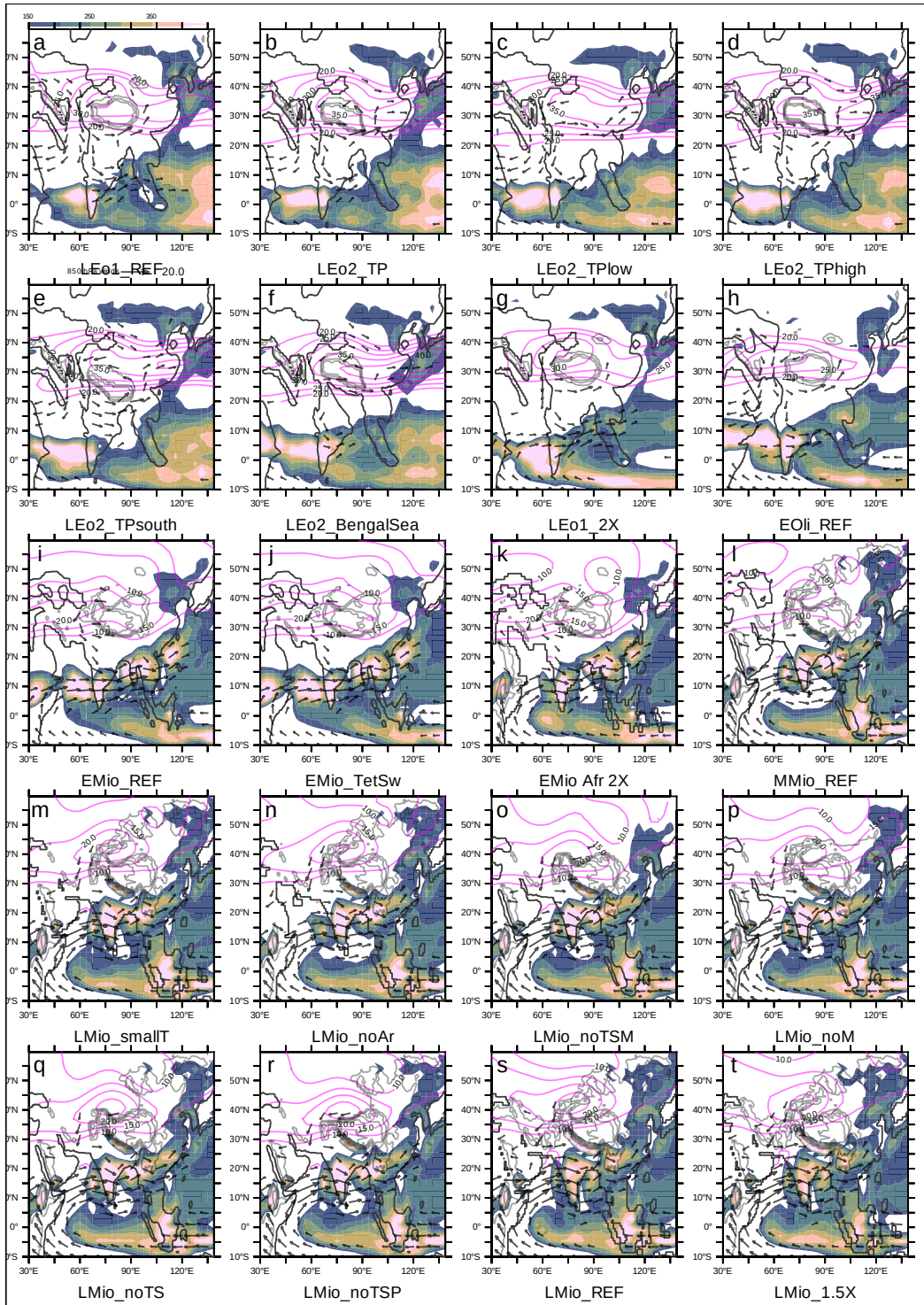


**Figure 17.** June-August (JJA) normalized Sea Level Pressure anomalies (shading, hPa) and 850 hPa winds over 4 m/s (vectors), for all reference and sensitivity experiments.

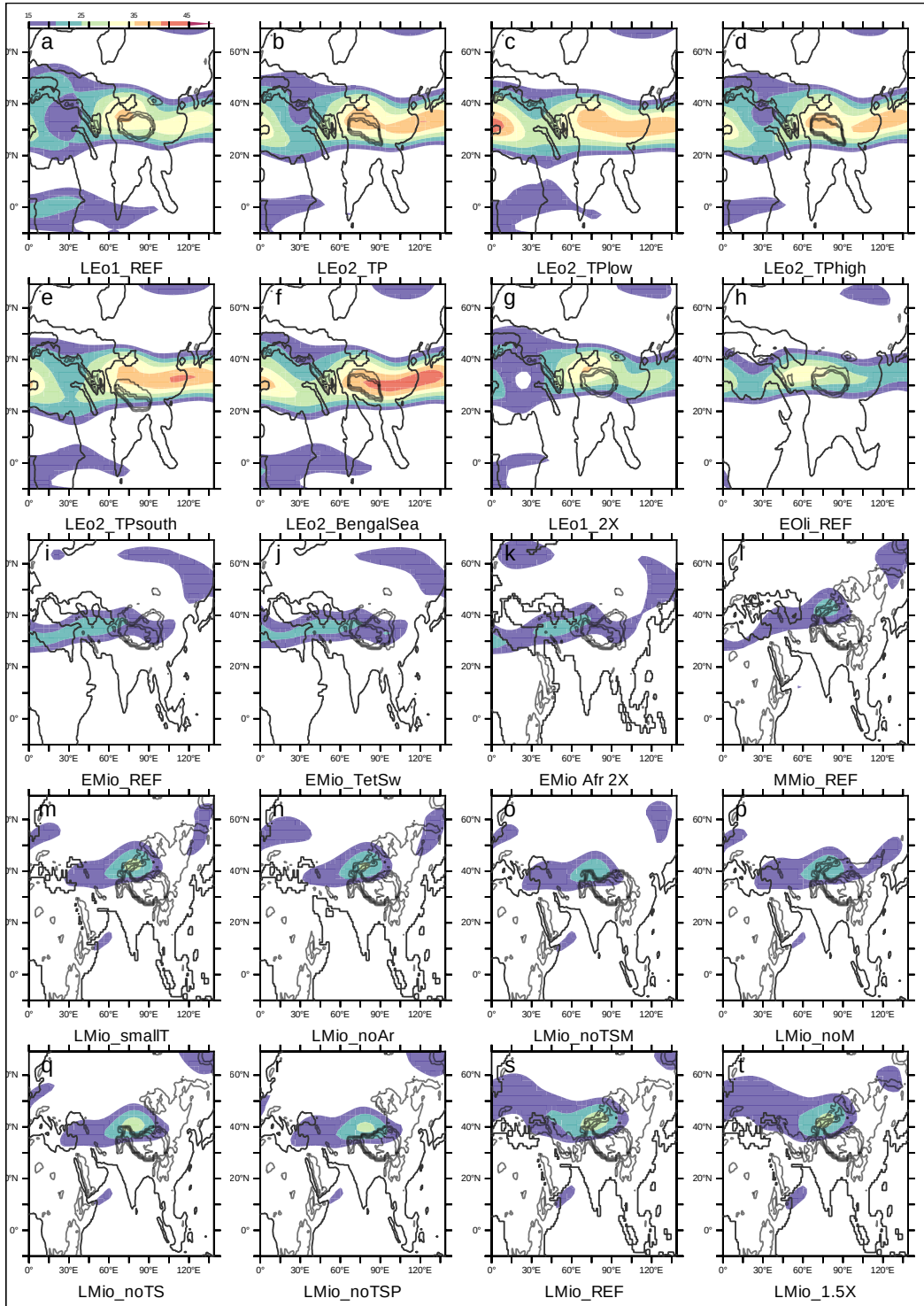


**Figure 18.** December-February (DJF) normalized Sea Level Pressure anomalies (shading, hPa) and 850 hPa winds over 4 m/s (vectors), for all reference and sensitivity experiments.



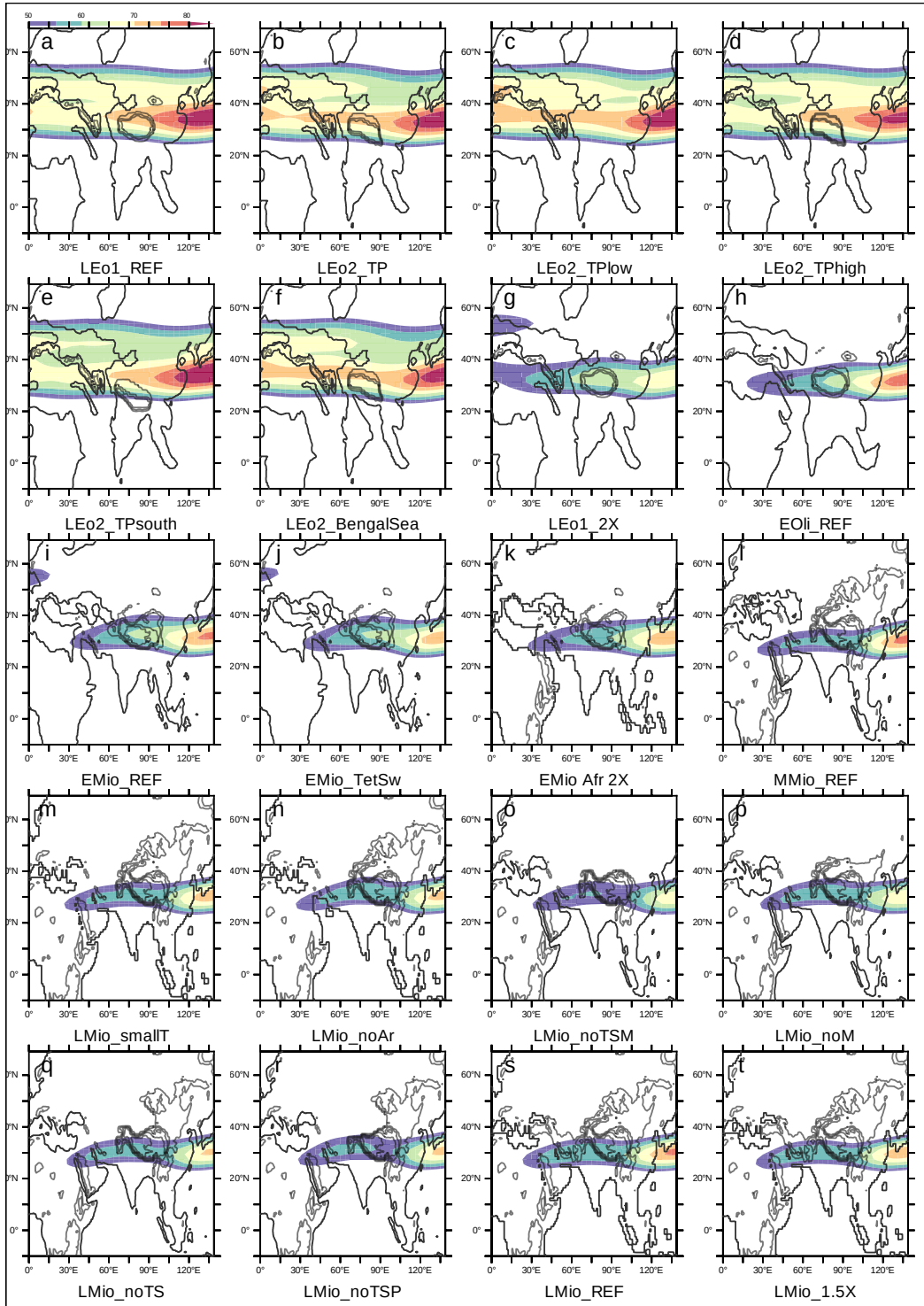


**Figure 19.** Jet Stream velocity and localization over Asia in August (purple outline), for all sensitivity experiments and precipitation for the month of August (shading, mm/month).



**Figure 20.** Jet Stream August velocity and localization over Asia, for all sensitivity experiments.





**Figure 21.** Jet Stream April velocity and localization over Asia, for all sensitivity experiments.