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Response of the Asian summer monsoons to idealized precession and obliquity forcing in a set of GCMs

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Abstract

1

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4

We examine the response of the Indian and East Asian summer monsoons to separate precession and obliquity forcing, using a set of fully coupled high-resolution models for the first time: EC-Earth, GFDL CM2.1,

5 CESM and HadCM3. We focus on the effect of insolation changes on
6 monsoon precipitation and underlying circulation changes, and find strong
7 model agreement despite a range of model physics, parameterization, and
8 resolution. Our results show increased summer monsoon precipitation
9 at times of increased summer insolation, i.e. minimum precession and
10 maximum obliquity, accompanied by a redistribution of precipitation and
11 convection from ocean to land. Southerly monsoon winds over East Asia
12 are strengthened as a consequence of an intensified land-sea pressure gra-
13 dient. The response of the Indian summer monsoon is less straightforward.
14 Over south-east Asia low surface pressure is less pronounced and winds
15 over the northern Indian Ocean are directed more westward. An Indian
16 Ocean Dipole pattern emerges, with increased precipitation and convec-
17 tion over the western Indian Ocean. Increased temperatures occur during
18 minimum precession over the Indian Ocean, but not during maximum
19 obliquity when insolation is reduced over the tropics and southern hemi-
20 sphere during northern hemisphere summer. Evaporation is reduced over
21 the northern Indian Ocean, which together with increased precipitation
22 over the western Indian Ocean dampens the increase of monsoonal precipi-
23 tation over the continent. The southern tropical Indian Ocean as well as
24 the western tropical Pacific (for precession) act as a moisture source for
25 enhanced monsoonal precipitation. The models are in closest agreement
26 for precession-induced changes, with more model spread for obliquity-
27 induced changes, possibly related to a smaller insolation forcing. Our re-
28 sults indicate that a direct response of the Indian and East Asian summer
29 monsoons to insolation forcing is possible, in line with speleothem records
30 but in contrast to what most marine proxy climate records suggest.

31 **Keywords:** monsoon, orbital forcing, paleoclimate modeling, South-East
32 Asia, multi-model, climate dynamics

33 1 Introduction

34 Monsoon systems play a key role in Asian climate, representing a strong sea-
35 sonal climate signal over an area spanning from the Arabian to the Chinese
36 Seas. The summer monsoon onset occurs in late spring / early summer for
37 the East Asian monsoon, and in summer for the Indian monsoon, when the
38 Intertropical Convergence Zone (ITCZ) rapidly moves northward towards the
39 continent, conveying large amounts of moisture and energy (e.g. Bordoni and
40 Schneider, 2008; Molnar et al., 2010; Mohtadi et al., 2016). On time scales of
41 10^3 - 10^5 years, the Asian monsoons are dominated by changes in the distribu-
42 tion of incoming solar radiation, the orbital or so-called Milankovitch cycles.
43 This cyclic variation in the spatial and temporal distribution of radiation has
44 a strong influence on Earth's climate (e.g. Ruddiman, 2006b; Mohtadi et al.,
45 2016). Precession controls the seasonality of insolation at all latitudes and is
46 modulated by the eccentricity of the earth's orbit, while obliquity (tilt) affects
47 mostly high latitude summer insolation and meridional insolation gradients. All
48 three orbital parameters (precession, eccentricity, and obliquity) are observed
49 in proxy climate records of monsoon strength. Examples of such records are
50 oxygen isotope speleothem records from east China (e.g. Wang et al., 2008) as
51 well as India (e.g. Kathayat et al., 2016), dominated by precession cyclicity, as
52 well as the multi-proxy stack of Indian summer monsoon circulation strength
53 from the western Arabian Sea, where southwesterly summer monsoon winds
54 influence upwelling, productivity and sedimentation (e.g. Clemens and Prell,
55 2003). The latter shows a strong obliquity signal as well, despite the dominance
56 of precession in low-latitude summer insolation.

57 Despite the remaining controversies in the interpretation of oxygen isotope
58 speleothem records (e.g. Caley et al., 2014; Mohtadi et al., 2016) the strong pre-
59 cession signal in phase with insolation in Chinese and Indian speleothem records
60 (e.g. Wang et al., 2008; Cai et al., 2015; Kathayat et al., 2016) is in line with cli-
61 mate model simulations (e.g. Battisti et al., 2014; Rachmayani et al., 2016). The

62 interpretation of the Arabian Sea proxies, originally thought to show a long lag
63 of monsoon strength with respect to precession (e.g. Clemens and Prell, 2003),
64 also remains an item of discussion (e.g. Ziegler et al., 2010; Caley et al., 2011),
65 with a recent modeling study suggesting that Arabian Sea productivity, part of
66 the multi-proxy stack, is not necessarily enhanced at times of a stronger Indian
67 Summer Monsoon (Le Mézo et al., 2016). This could explain the discrepancy in
68 lags between proxy studies and speleothem records as well as modeling studies,
69 with the latter showing no lags. Modelling studies corroborate the strength-
70 ening of summer monsoons at times of orbitally forced high summer insolation
71 and the weakening at times of orbitally forced weak insolation, even if experi-
72 ments are run for only up to a few hundred years (i.e. short on the orbital time
73 scale). In some of the earliest paleoclimate modelling studies, atmosphere-only
74 models showed a strengthened thermal low over the continents and a stronger
75 land/sea thermal contrast, causing increased summer monsoon precipitation
76 at times of high summer insolation (e.g. Kutzbach and Otto-Bliesner, 1982;
77 Kutzbach and Guetter, 1986; Prell and Kutzbach, 1987). More recently several
78 studies of the Mid-Holocene, a time of enhanced Northern Hemisphere inso-
79 lation seasonality, were performed within the framework of the Paleoclimate
80 Modelling Intercomparison Project (PMIP, Braconnot et al. (2007)). During
81 the Mid-Holocene, models show a stronger Indian Summer Monsoon (ISM) and
82 East Asian Summer Monsoon (EASM). The EASM strengthening is related to
83 a stronger land/sea pressure gradient (Jiang et al., 2013; Wang and Wang, 2013;
84 Zheng et al., 2013). Strengthening of the ISM may be affected by mechanisms
85 such as the Indian Ocean Dipole (Zhao et al., 2005; Abram et al., 2007). Other
86 studies of periods with a precession-induced increase in insolation seasonality
87 have also demonstrated a strengthening of the ISM (Braconnot and Marti, 2003;
88 Braconnot et al., 2008; Battisti et al., 2014; Araya-Melo et al., 2015; Rachmayani
89 et al., 2016).

90 Only a few studies have investigated the separate precession and obliquity
91 forcing instead of focusing on a specific time with combined precession and obliq-

92 uity forcing. Tuentner et al. (2003) showed a deepening of the convergence zone
93 over southern Asia and increased summer precipitation over the Asian mon-
94 soon regions during both minimum precession and maximum obliquity (both
95 times of increased summer insolation). However, we have already shown that
96 the mechanisms behind the response to orbital forcing in their model is rather
97 different from the response in the EC-Earth model used here, specifically for
98 the North African monsoon Bosmans et al. (2015a). Nonetheless, the orbitally
99 induced changes in precipitation are similar to those identified by Erb et al.
100 (2013). Mantsis et al. (2013) report increased precipitation during minimum
101 precession as well, which for East Asia is related to reduced pressure over land
102 as well as an increased North Pacific high pressure area, both intensifying the
103 land/sea pressure gradient. This is also modeled by Shi et al. (2011) for both
104 precession and obliquity and by Wang et al. (2012) for precession. Chen et al.
105 (2011b) focus solely on obliquity, showing that the ISM and the South-EASM
106 are stronger during maximum obliquity, while the North-EASM is weaker. Wu
107 et al. (2016) show that high obliquity during the early Holocene augments the
108 impact of precession by affecting high pressure systems and meridional gradi-
109 ents in pressure and temperature. Multiple studies have found that the orbital-
110 induced changes in surface pressure over the South Asian monsoon regions do
111 not show a straightforward change in land/sea pressure differences (Zhao et al.,
112 2005; Chen et al., 2011b; Mantsis et al., 2013).

113 The link between orbitally forced changes in insolation and monsoon strength
114 has thus been established by both proxy climate records and modelling studies.
115 Here, we focus on the mechanisms behind changes in summer monsoon strength
116 using state-of-the-art general circulation models, assessing the detailed pattern
117 of the ISM and EASM response to both precession and obliquity forcing using
118 fully coupled general circulation models (EC-Earth, GFDL, CESM for preces-
119 sion and obliquity as well as HadCM3 for obliquity). These models cover a range
120 of model physics, parameterization and resolution. Such a multi-model approach
121 provides the opportunity to judge whether results are model-dependent, and if

122 this is not the case, provides robust mechanisms behind the orbital signals ob-
123 served in proxy records. We single out the effects of precession and obliquity,
124 as the latter has a relatively strong impact on monsoon strength given its weak
125 impact on low-latitude insolation (e.g. Tuenter et al., 2003). Using idealized
126 experiments enables us to separate and maximize the precession and obliquity
127 signals in our experiments.

128 This paper is organised as follows: Section 2 describes each of the general
129 circulation models and the experimental set-up. Section 3 shows the changes in
130 monsoon precipitation and associated circulation, with Section 3.1 focusing on
131 precession and Section 3.2 focussing on obliquity. A discussion and conclusion
132 are given in Sections 4 and 5.

133 **2 Model and Experiment set-up**

134 **2.1 EC-Earth**

135 EC-Earth is a fully coupled ocean-atmosphere GCM (general circulation model,
136 Hazeleger et al. 2010, 2011). The atmospheric part of EC-Earth 2.2 is based on
137 the Integrated Forecasting System (IFS), cycle 31R1, of the European Centre for
138 Medium-range Weather Forecast (ECMWF). Its horizontal resolution is T159
139 (roughly $1.125^\circ \times 1.125^\circ$) with 62 vertical levels. The ocean model NEMO runs
140 at a resolution of nominally 1° with 42 vertical levels. The ocean, ice, land and
141 atmosphere are coupled through the OASIS3 coupler (Valcke and Morel, 2006).

142 EC-Earth has previously been shown to represent monsoons well in both the
143 pre-industrial and the Mid-Holocene paleo-experiment (Bosmans et al., 2012).
144 Furthermore, the orbital extreme experiments used in this paper were also used
145 to investigate orbital forcing of the North-African monsoon (Bosmans et al.,
146 2015a).

147 **2.2 GFDL CM2.1**

148 The GFDL CM2.1 model is the Geophysical Fluid Dynamics Laboratory Cli-
149 mate Model version 2.1, (Delworth et al., 2006). Like EC-Earth 2.2 this is an
150 ocean-atmosphere fully coupled model, including land and sea ice components.
151 It runs at a resolution of 2° latitude by 2.5° longitude with 24 vertical levels
152 and an ocean resolution of $1^\circ \times 1^\circ$, with higher meridional resolution near the
153 equator, and 50 vertical levels. The atmospheric model has a time step of 3
154 hours for radiation and 30 min for other atmospheric physics.

155 GFDL-CM2.1 has previously been used to investigate climatic response to
156 orbital forcing (e.g. Mantsis et al., 2013). The orbital experiments used here
157 are the same as in Erb et al. (2015), where the climatic response to changes in
158 obliquity, precession, CO_2 and ice sheets is investigated. Here we use the orbital
159 experiments with pre-industrial greenhouse gas concentrations and ice sheets.

160 **2.3 CESM**

161 The GFDL experiments were repeated with the National Center for Atmospheric
162 Research’s (NCAR) Community Earth System Model 1.2 (CESM1.2), which is
163 also a fully coupled atmosphere-ocean model. CESM1.2 includes the CAM5
164 Community Atmospheric Model at 2.5×1.875 resolution with a 30 minute time
165 step and 30 vertical levels, the POP2 Parallel Ocean Program as the oceanic
166 component, running at approximately 1×0.5 resolution with 60 vertical levels,
167 and the Community Land Model CLM4.0. Fixed vegetation is used. Here we
168 use the same idealized simulations previously used to study the climate response
169 to changes in obliquity and other past forcings (Erb et al., 2018).

170 **2.4 HadCM3**

171 HadCM3 is the UK Met Office Hadley Centre Coupled Climate Model Version3.
172 Its horizontal resolution of the atmosphere model is 2.5° in latitude by 3.75° in
173 longitude and consists of 19 layers in the vertical, comparable to a T42 spectral

174 model resolution. The atmospheric model has a time step of 30 min. The spatial
175 resolution over the ocean is $1.25^\circ \times 1.25^\circ$ with 20 vertical layers. The sea-ice
176 model uses a simple thermodynamic scheme and contains parameterisations of
177 ice drift and leads (Cattle et al., 1995).

178 HadCM3 is well documented (Gordon et al., 2000) and has previously been
179 shown to reproduce the main features of modern climate observations. Further-
180 more, HadCM3 has been used in the past to examine the effect of orbital forcing
181 in the Quaternary (e.g. Singarayer and Valdes, 2010) and in earlier periods such
182 as the mid-Pliocene (e.g. Dolan et al., 2011; Prescott et al., 2014).

183 **2.5 Experimental set-up: insolation forcing and boundary** 184 **conditions**

185 This study is based on experiments of orbital extremes, with EC-Earth, GFDL-
186 CM2.1 and CESM running both precession and obliquity extremes and HadCM3
187 running the obliquity extremes. These model simulations form an ensemble
188 of opportunity rather than being part of a pre-defined model intercomparison
189 project. As a result, there are small differences in the experimental design. The
190 main differences between the experiments in all models are the orbital param-
191 eters, and thus the insolation forcing, but there are small differences in the exact
192 orbital parameters and greenhouse gas concentrations. Generally, EC-Earth
193 and HadCM3 have the same set-up, as do GFDL-CM2.1 and CESM. Table 1
194 shows the set-up per experiment and per model. Four time-slice experiments
195 are performed to examine the separate precession and obliquity signals: min-
196 imum and maximum precession (Pmin and Pmax) as well as maximum and
197 minimum obliquity (Tmax and Tmin, T for tilt), allowing us to maximize the
198 orbital signals from our experiments. All simulations are performed with fixed
199 present-day ice sheets and vegetation.

200 During a precession minimum (Pmin) the summer solstice (midsummer)
201 occurs at perihelion (the point in the earth’s orbit closest to the Sun), so sea-
202 sonality is enhanced on the Northern Hemisphere and reduced on the Southern

Table 1: Overview of the orbital configuration in each experiment. Obl is the obliquity (tilt, in degrees), $\tilde{\omega}$ is the longitude of perihelion, defined here as the angle from the vernal equinox to perihelion in degrees, measured counterclockwise. e is eccentricity. $e \sin(\pi+\tilde{\omega})$ is the precession parameter. Note that for Tmax and Tmin there is no precession when a circular orbit ($e=0$) is used. For GHGs the year of the greenhouse gas concentrations is given, with the CO₂ concentration in parentheses in ppmv. Calendar anchor point is either vernal equinox (v.e.), autumnal equinox (a.e.) or not applicable (n.a.) when $e=0$. An asterix (*) indicates that the model output has been processed onto a fixed-angular calendar.

	Obl (°)	$\tilde{\omega}$ (°)	e	$e \sin(\pi+\tilde{\omega})$	GHGs	Calendar anchor point
Pmax (perihelion at NH winter)						
EC-Earth	22.08	273.50	0.058	0.058	1850 (284.5)	v.e.
GFDL-CM2.1	23.439	270	0.0493	0.0493	1860 (286)	a.e.*
CESM	23.439	270	0.0493	0.0493	1860 (286)	v.e.*
Pmin (perihelion at NH summer)						
EC-Earth	22.08	95.96	0.056	-0.055	1850 (284.5)	v.e.
GFDL-CM2.1	23.439	90	0.0493	-0.0493	1860 (286)	a.e.*
CESM	23.439	90	0.0493	-0.0493	1860 (286)	v.e.*
Tmax (maximum obliquity)						
HadCM3	24.45	-	0	0	1850 (284.5)	n.a.
EC-Earth	24.45	-	0	0	1850 (284.5)	n.a.
GFDL-CM2.1	24.480	282.93	0.0167	0.0163	1860 (286)	a.e.
CESM	24.480	282.93	0.0167	0.0163	1860 (286)	v.e.
Tmin (minimum obliquity)						
HadCM3	22.08	-	0	0	1850 (284.5)	n.a.
EC-Earth	22.08	-	0	0	1850 (284.5)	n.a.
GFDL-CM2.1	22.079	282.93	0.0167	0.0163	1860 (286)	a.e.
CESM	22.079	282.93	0.0167	0.0163	1860 (286)	v.e.

203 Hemisphere. The opposite occurs during a precession maximum (Pmax), when
204 winter solstice occurs at perihelion. In the obliquity experiments, eccentricity
205 is set to zero to completely eliminate the effect of precession in EC-Earth and
206 HadCM3, a small value of eccentricity is used in GFDL-CM2.1 and CESM.
207 During an obliquity maximum (Tmax, T for tilt), both northern and south-
208 ern hemisphere (NH, SH) summers receive more insolation, especially at the
209 poles, while during an obliquity minimum (Tmin) summer insolation is reduced.
210 Within one season, precession has the same effects on both hemispheres, while
211 obliquity has the opposite effect. The values of the orbital parameters in each
212 experiment are given in Table 1. For EC-Earth and HadCM3 these are the same
213 as the P-T-, P+T-, P0T+, P0T- experiments in Tuentler et al. (2003), and are
214 based on the most extreme values of the orbital parameters occurring in the last
215 1 Ma (Berger, 1978).

216 Insolation differences at $\sim 40^\circ\text{N}$ can be as large as 100 Wm^{-2} for precession
217 and 20 Wm^{-2} for obliquity (Figure 1). Note that the insolation change between
218 the orbital extremes vary amongst the models due to slight differences in the
219 orbital parameters, as well as the choice of calendar. For experiments in which
220 eccentricity is not set to zero, the way the calendar is implemented can result
221 in changes in the timing of the equinoxes and solstices, which may affect model
222 results. In the EC-Earth precession experiments the vernal equinox is fixed at
223 March 21st and the present-day calendar is used. The same applies to CESM,
224 while GFDL-CM2.1 fixes the autumnal equinox at September 21st. Both the
225 CESM and GFDL-CM2.1 monthly output is then corrected to fixed-angular
226 “months” following Pollard and Reusch (2002) in order to account for this cal-
227 endar effect (Erb et al., 2015). Figure A.1 shows the difference in the insolation
228 changes. Studies have found that the calendar-effect has only a minor effect on
229 the results (e.g. Chen et al., 2011a), also in HadCM3 seasonal results (Marzocchi
230 et al., 2015). Here, we also find that for CESM and GFDL the results shown
231 in this paper are not changed by the choice of calendar. Only the annual cycle
232 changes slightly, but the patterns of change in summer that we focus on here

233 remain the same, see Figures A.2, A.3, A.4. Despite the small differences in
234 imposed forcings and calendars, we find that monsoonal responses are robust
235 amongst models, further suggesting that the results are not overly sensitive to
236 the exact experimental design.

237 In this study we compare Pmin to Pmax, and Tmax to Tmin, i.e. we
238 investigate the effect of increased summer and decreased winter insolation on the
239 Northern Hemisphere. EC-Earth experiments were run for 100 years, of which
240 the last 50 years are used to create the climatologies shown in this study. This
241 is long enough for top-of-atmosphere net radiation as well as atmospheric and
242 surface variables that are of interest to equilibrate to the forcing (see Bosmans
243 et al. (2015a)). The globally averaged tendency term of surface air temperature,
244 dT/dt , is near-zero and shows no trend in all experiments (not shown). HadCM3
245 was run for 300 years per experiment, of which the last 50 years are used.
246 GFDL-CM2.1 and CESM were run for at least 600 and 500 years respectively
247 and 100-year climatologies were computed.

248 **3 Results**

249 In this section we first investigate the precession-induced changes in the Asian
250 monsoons (Section 3.1), followed by the obliquity-induced changes (Section 3.2).
251 We compare maximum to minimum NH summer insolation, i.e. Pmin to Pmax
252 and Tmax to Tmin, using JJA averages. Precipitation results are shown for all
253 models, other variables are shown for EC-Earth only for brevity. Results of all
254 models are shown in the supplementary material (Section C).

255 **3.1 Precession**

256 Within the experiments presented here, the precession-induced insolation change
257 reaches 100 Wm^{-2} in June (Figure 1) (Tuenter et al., 2003; Bosmans et al.,
258 2015a). The JJA averaged insolation between 10°N and 40°N is $\sim 80 \text{ Wm}^{-2}$
259 higher during Pmin than Pmax. Figure 2 shows that the average summer precip-

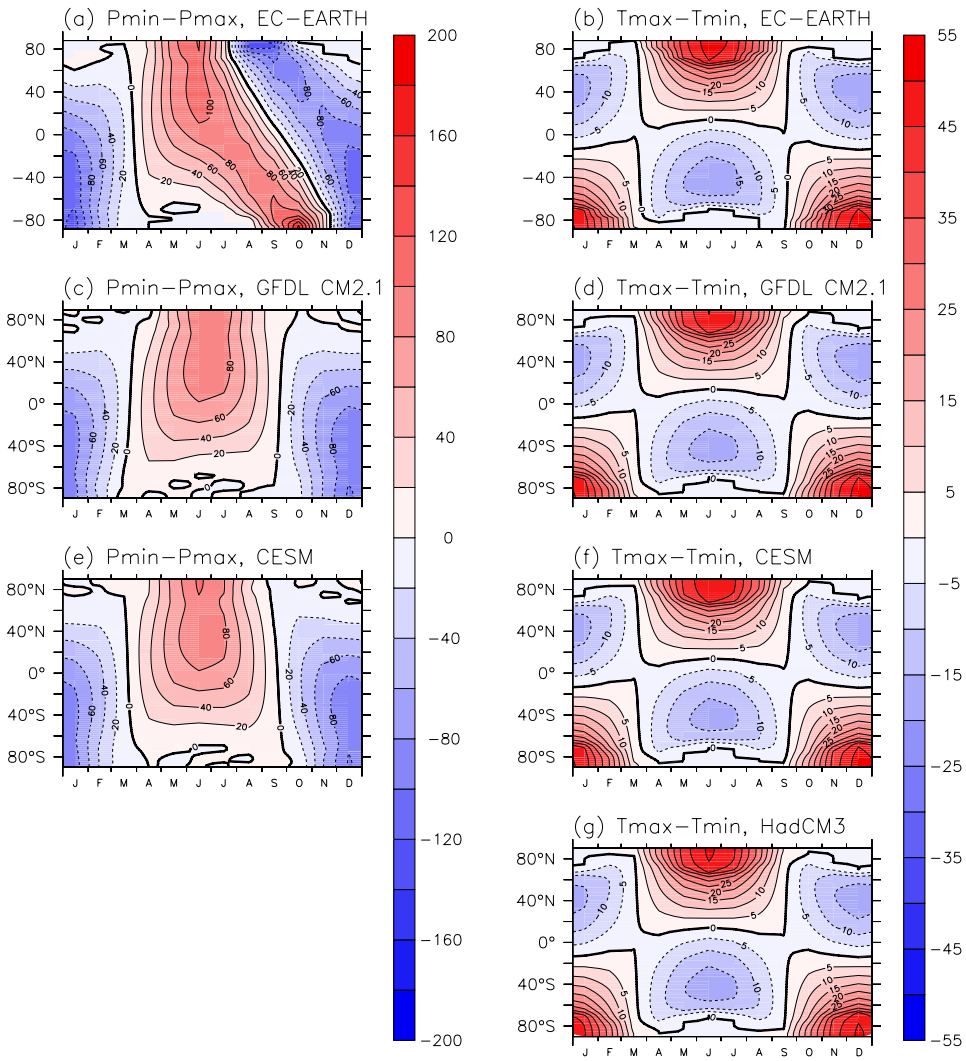


Figure 1: Insolation changes in W/m^2 per model and for the precession and obliquity experiments. See Table 1 for details on the orbital configuration per experiment. Note that output of the CESM and GFDL precession experiments has been processed onto a fixed-angular calendar, explaining the difference in precession-induced insolation change compared to EC-Earth, whose output remained on the fixed-day calendar used in the experiment. The range of insolation difference for precession (up to $\sim 100 \text{ W/m}^2$) is much larger than for obliquity (up to $\sim 50 \text{ W/m}^2$).

260 itation over monsoonal Asia is up to 3 mm/day higher during Pmin in EC-Earth,
261 up to 2 mm/day higher in GFDL and up to 2.5 mm/day higher in CESM. Fur-
262 thermore, in line with the insolation forcing, the seasonality is greater in Pmin.
263 The largest precipitation changes occur over the Himalaya, just south of the
264 Tibetan plateau, see Figure 3. Models are also consistent in producing more
265 precipitation during Pmin over most of the South-East Asian Peninsula, Indone-
266 sia and the western Indian Ocean. Reduced precipitation occurs over the eastern
267 Indian Ocean, Bay of Bengal and the Chinese Seas. East of the Tibetan Plateau
268 CESM simulates reduced precipitation as well, whereas EC-Earth shows slightly
269 more precipitation, as does GFDL. This could be related to CESM having much
270 more precipitation in the Pmax experiment in this area than the other models
271 (contours in Figure 3). Models differ over India as well, with CESM and EC-
272 Earth for instance showing high precipitation just west of the Western Ghats
273 during Pmax, and lower precipitation during Pmin. This could be related to
274 representation of orography (Figure B.1).

275 To assess the precipitation changes in more detail, we first investigate changes
276 in surface temperature, surface pressure and surface winds. The hydrological
277 cycle and upper level circulation features will be discussed in later paragraphs.

278 For precession, higher summer insolation results in higher surface air tem-
279 peratures (Figure 4), except for monsoonal North-Africa / westernmost Arabian
280 Peninsula and northwest India / Pakistan. Strong increases in cloud cover over
281 these areas (not shown) decrease the amount of solar radiation reaching the
282 surface. In addition, increased evaporation cools the surface. These monsoon-
283 intensification feedbacks thus completely overcome the direct warming effect of
284 increased insolation. In CESM these feedbacks seem particularly strong, re-
285 sulting in a stronger cooling over a larger area of India and Pakistan than in
286 EC-Earth and GFDL (Figure C.1). The rest of the continent warms up strongly,
287 more than 8°C over continental Asia and 10°C over the Middle East. Warming
288 over the ocean is smaller due to its large heat capacity. Over south-east Asia,
289 the temperature response over land (south of $\sim 25^\circ\text{N}$) is dampened by a small

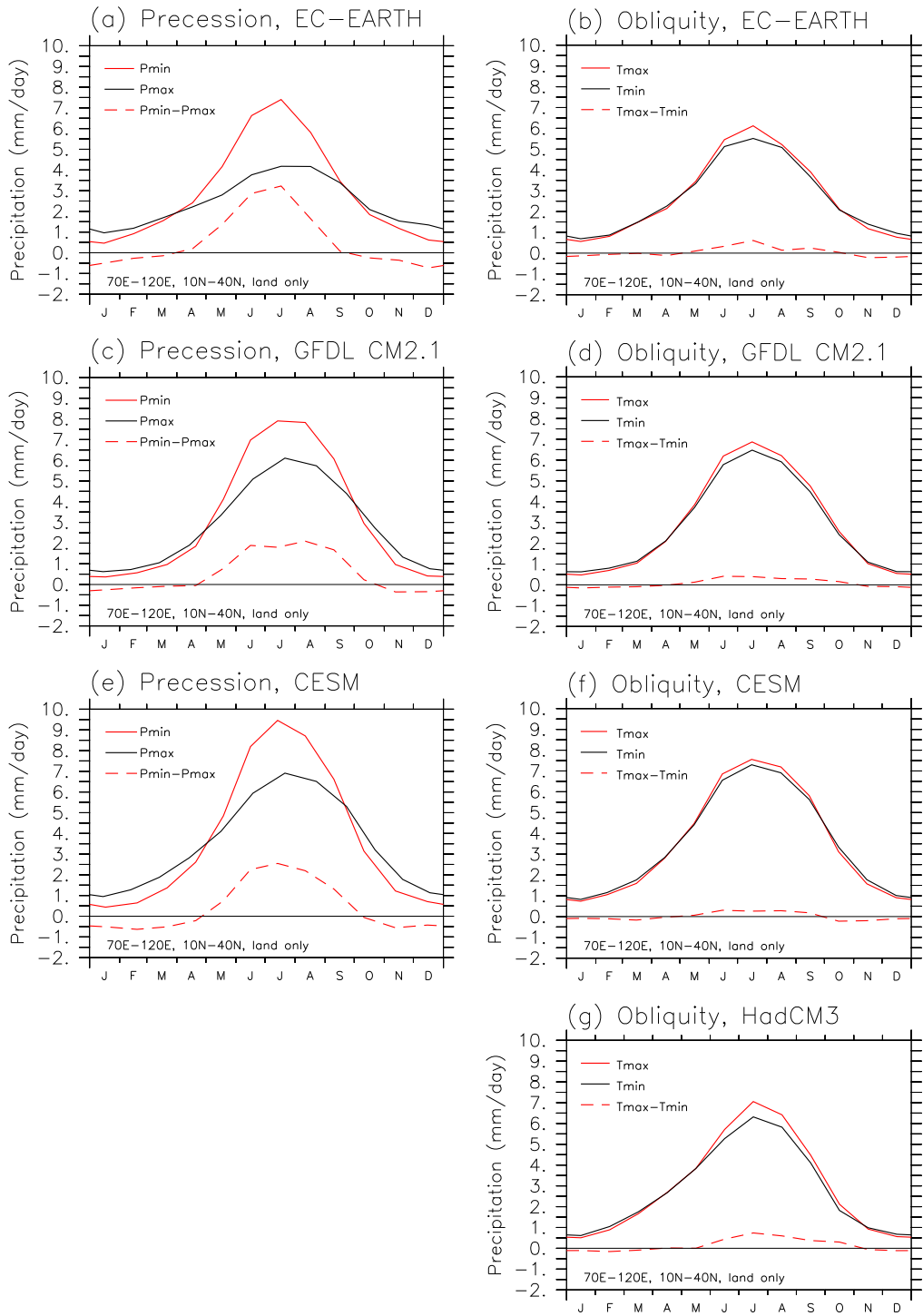


Figure 2: Precipitation over Asia per model, in mm/day, averaged over 70°E:120°E, 10°N:40°N land only for precession (a,c,e) and obliquity (b,d,f,g). Differences are given by the dashed lines.

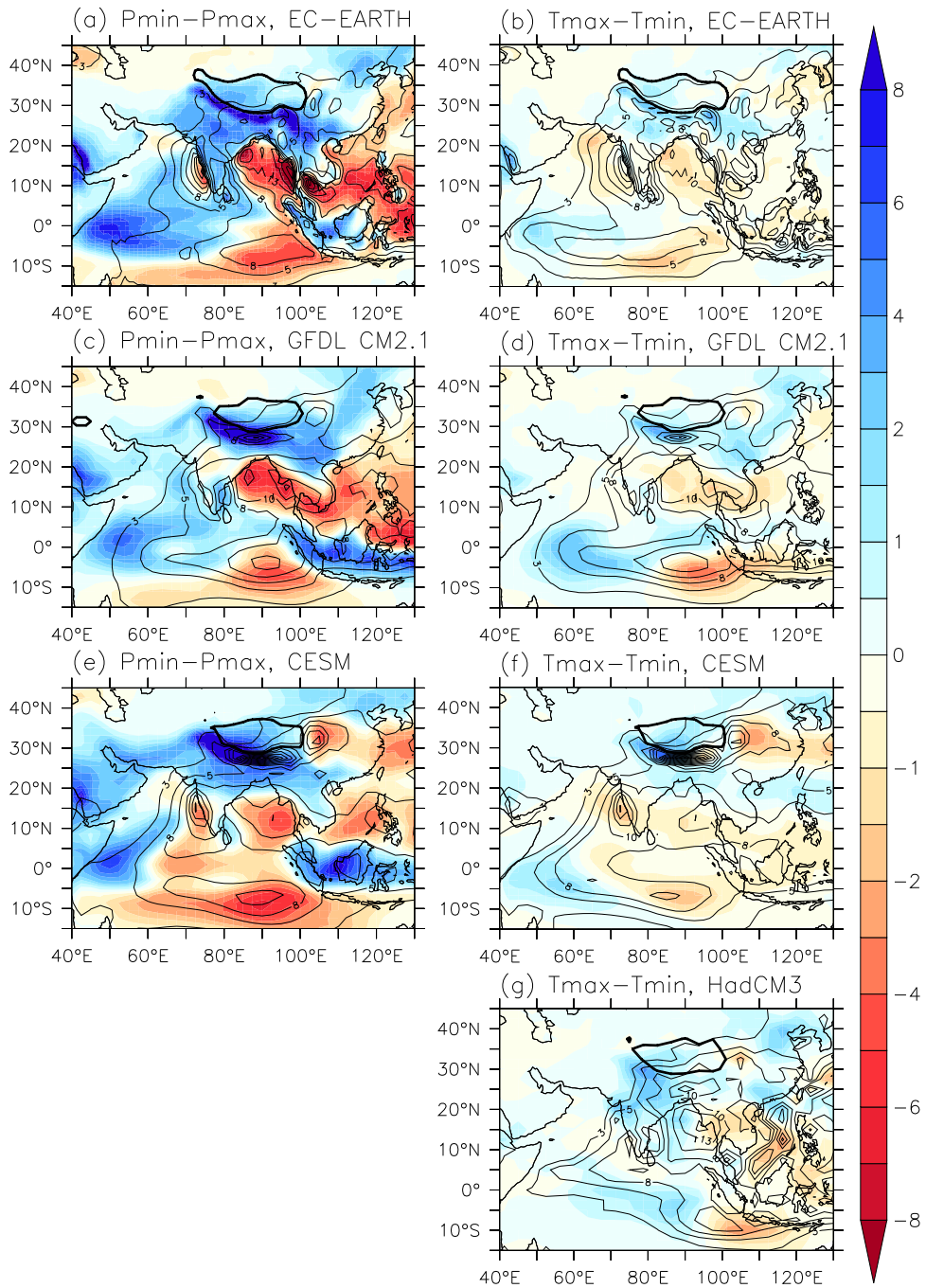


Figure 3: Difference in June-July-August average precipitation in mm/day for Pmin-Pmax (left) and Tmax-Tmin (right) per model. Contours indicate values for Pmax (left; a, c, e) or Tmin (right; b, d, f, g). The thick contour line is at 4km height, indicating the Tibetan Plateau.

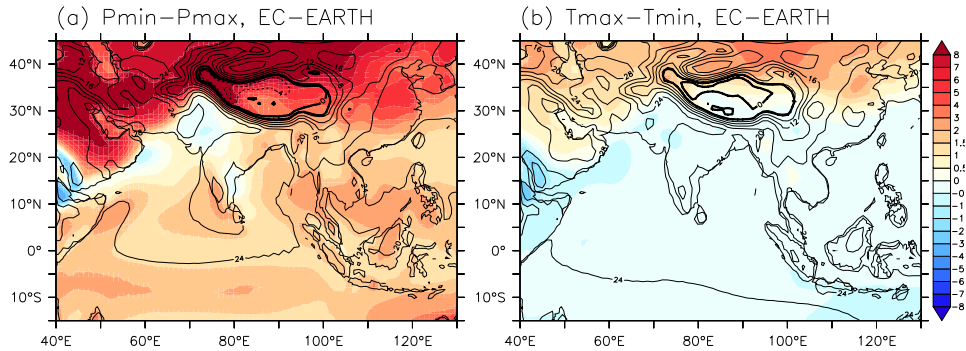


Figure 4: June-July-August average surface air temperature difference for Pmin-Pmax (left) and Tmax-Tmin (right) for EC-Earth. Results for all models can be found in the Supplementary Materials (Figure C.1). Contours indicate values for Pmax (left) or Tmin (right) in $^{\circ}\text{C}$. The thick contour line is at 4km height, indicating the Tibetan Plateau.

290 increase in cloud cover and increased evaporative cooling.

291 In response to increased summer temperatures over the continent, sea level
 292 pressure over these regions is reduced (Figure 5), mostly over continental Asia
 293 and the Middle East. Over the Tibetan Plateau, southern India, the Bay of
 294 Bengal, South-East Asia and the Chinese Seas, sea level pressure is higher during
 295 Pmin in EC-Earth. The area of higher surface pressure over South-East Asia is
 296 connected to a strengthened North Pacific High (Figure 5) in all models (Figures
 297 C.2, C.3). CESM displays a stronger pressure increase over southern and south-
 298 east Asia as well as over the Indian Ocean compared to EC-Earth and GFDL
 299 (Figure C.3).

300 The strengthened North Pacific High and the lower surface pressure over
 301 central and eastern Asia force stronger southerly moisture transport over the
 302 EASM (Figure 5), related to stronger southerly winds (Figure 7). Over the
 303 northern Indian Ocean, the high pressure anomaly pushes winds and moisture
 304 transport more westward and reduces windspeed through a weaker meridional
 305 pressure gradient between the equator and $\sim 10\text{-}15^{\circ}\text{N}$. Just south of the equator
 306 the meridional pressure gradient is stronger and winds as well as moisture trans-
 307 port are stronger (Figures 5, 7). Monsoonal winds and moisture transport over
 308 the northernmost Arabian Sea and Bay of Bengal are stronger as well. Wind

309 and moisture transport changes in GFDL and CESM are very similar, except
310 over the northernmost Arabian Sea / south-eastern Arabian peninsula (Figures
311 C.7, C.2, C.3).

312 To investigate the source of the increased monsoon precipitation during
313 Pmin, we also considered evaporation. Figure 8 shows that evaporation over
314 land is increased in most areas. This increase, up to 3 mm/day, is small com-
315 pared to the precipitation increase, which reaches 15 mm/day in EC-Earth
316 (Figure 3). Precipitation is redistributed with less precipitation over the sur-
317 rounding oceans (except the western tropical Indian Ocean) and more over land
318 during Pmin (Figure 3). There is no additional moisture source from ocean
319 evaporation over the northern Indian Ocean (Figure 8), where evaporation is
320 reduced in relation to reduced wind speed (Figure 7). Just south of the equator
321 evaporation and wind speed are higher during Pmin, so this southern hemi-
322 sphere region can act as a source of enhanced precipitation over the western
323 Indian Ocean as well as the northern hemisphere (NH) Asian continent. Fur-
324 thermore, looking at a larger area reveals enhanced moisture transport from
325 the western tropical Pacific into the Indian Ocean. As a result of enhanced
326 surface pressure over both the North and South Pacific (Figure 5), westwards
327 wind and moisture transport (Figures 7 and 5) is strengthened at tropical lat-
328 itudes, extending westward moisture transport into the western Indian Ocean.
329 Over the western tropical Pacific, this results in lower net precipitation (Figure
330 5). The surface latent heat flux over regions of enhanced evaporation (south-
331 ern hemisphere tropical Indian Ocean, western tropical Pacific) is enhanced,
332 following the same patterns as evaporation, Figure 8. GFDL and CESM also
333 show an overall increase of evaporation over land as well as the southern Indian
334 Ocean (Figure C.8), and furthermore also display enhanced wind and moisture
335 transport from the western tropical Pacific into the Indian Ocean (Figures C.2,
336 C.3).

337 The enhanced precipitation and moisture transport into the ISM area de-
338 spite lower evaporation over the northern Indian Ocean, can thus be related

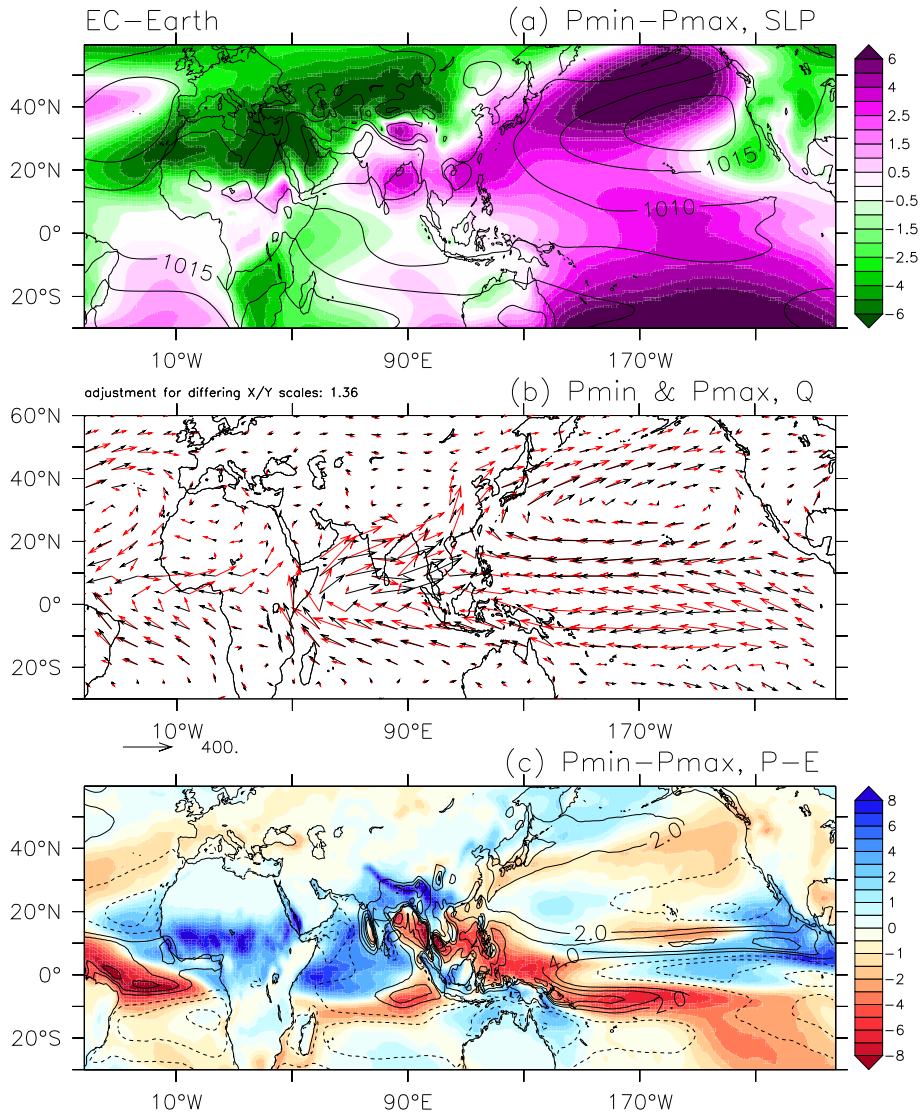


Figure 5: June-July-August average results for EC-Earth Pmin-Pmax. Top (a) shows sea level pressure difference in hPa with Pmax values in contours. Middle (b) shows moisture transport \mathbf{Q} , the vertical integral of $q\mathbf{v}$ in kg/(ms), during Pmin in red, and Pmax in black. Daily q and \mathbf{v} output from EC-Earth was used to compute \mathbf{Q} . Bottom (c) shows net precipitation with positive values (blue) indicating increased net precipitation in mm/day and contours showing net precipitation during Pmax JJA. Results for all models can be found in the Supplementary Materials (Figure C.2 and C.3).

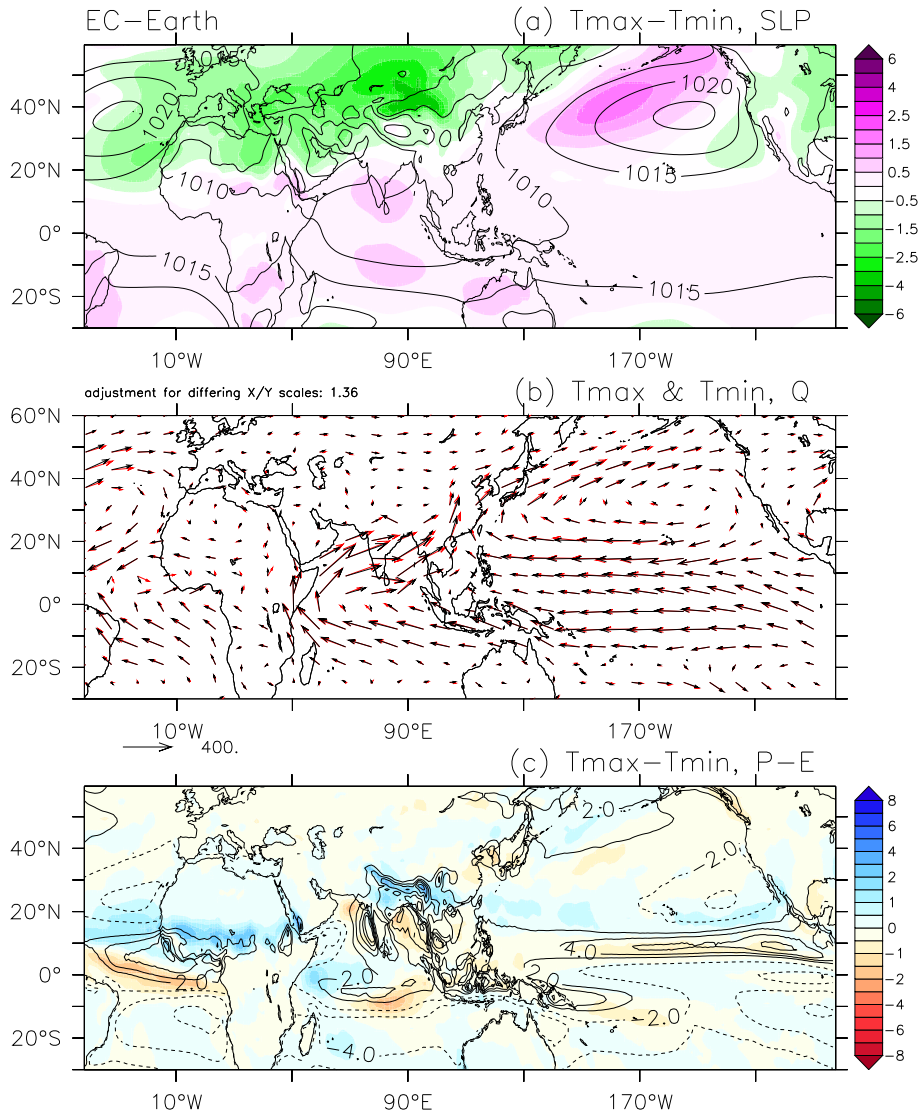


Figure 6: June-July-August average results for EC-Earth Tmax-Tmin. Top (a) shows sea level pressure difference in hPa with Tmin values in contours. Middle (b) shows moisture transport Q , the vertical integral of qv in $kg/(ms)$, during Tmax in red, and Tmin in black. Daily q and v output from EC-Earth was used to compute Q . Bottom (c) shows net precipitation with positive values (blue) indicating increased net precipitation in mm/day and contours showing net precipitation during Tmin JJA. Results for all models can be found in the Supplementary Materials (Figures C.4, C.5, C.6).

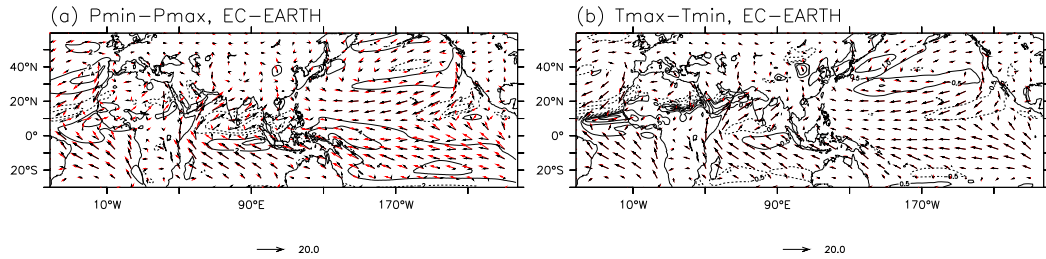


Figure 7: June-July-August average surface wind in m/s during Pmin (red) and Pmax (black, left), Tmax (green) and Tmin (blue, right) in EC-Earth. Contours indicate windspeed differences. Positive values are given by solid lines, negative values by dashed lines. The contour interval for precession (left) is 2 m/s and 0.5 m/s for obliquity (right). Results for all models can be found in the Supplementary Materials (Figure C.7).

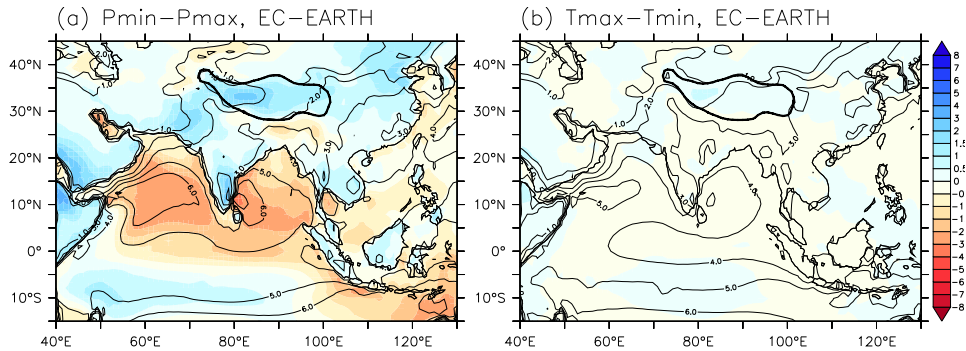


Figure 8: June-July-August average evaporation difference in mm/day for Pmin-Pmax (left) and Tmax-Tmin (right) for EC-Earth. Positive values (blue) indicate increased evaporation. Results for all models can be found in the Supplementary Materials (Figure C.8). Contours indicate values for Pmax (left) or Tmin (right). The thick contour line is at 4km height, indicating the Tibetan Plateau.

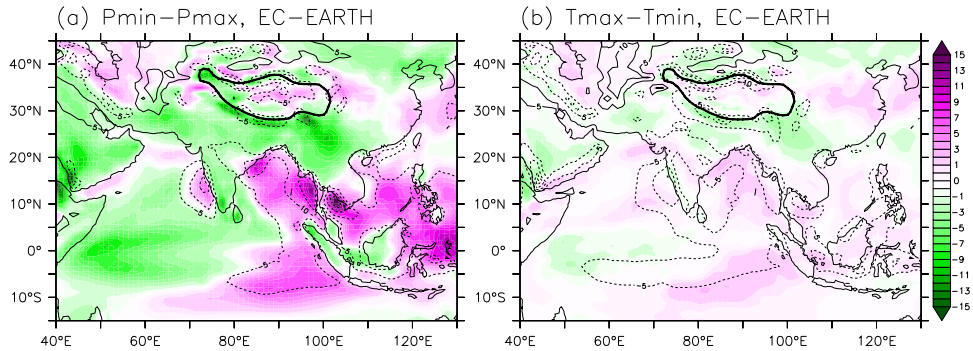


Figure 9: June-July-August average vertical velocity at 500 hPa in 10^{-2} Pa/s difference for Pmin-Pmax (left) and Tmax-Tmin (right). Contours indicate values for Pmax (left) or Tmin (right) with negative values indicating upward motion. Green indicates more upward or less downward motion, purple indicates more downward or less upward motion. Results for all models can be found in the Supplementary Materials (Figure C.13). The thick contour line is at 4km height, indicating the Tibetan Plateau.

339 to enhanced moisture transport from the southern Indian Ocean as well as the
 340 western tropical Pacific Ocean. Both the enhanced westward wind and moisture
 341 transport from the Pacific, as well as the reduced wind speeds over the northern
 342 Indian Ocean causing lower evaporation, are associated with anomalously high
 343 pressure. Increased specific humidity (not shown) over the northern Arabian Sea
 344 and East Asia plays a small role, but the major factor in the moisture transport
 345 changes is wind (compare Figure 7 and 5). A breakdown of moisture transport
 346 transport confirms the major role of wind in precession-induced moisture transport
 347 changes (see Supplementary Figure C.9). In CESM and GFDL the dynamic
 348 (wind-driven) part of moisture transport changes is strongest as well (Figures
 349 C.10, C.11). See Equation 1 in Bosmans et al. (2015a) for the breakdown of
 350 moisture transport into wind- and / or humidity-driven parts.

351 Changes in the middle troposphere are consistent with the surface precip-
 352 itation changes. Figure 9 shows stronger convection (upward motion) along
 353 the Himalayas during Pmin, as well as stronger convection over the rest of
 354 monsoonal Asia and the western Indian Ocean. Over the ocean regions where
 355 precipitation is lower, convection is reduced. The same holds for GFDL and
 356 CESM (Figure C.13).

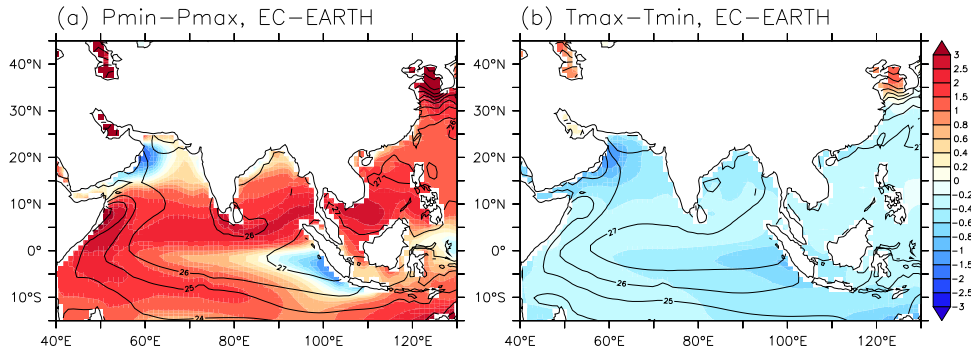


Figure 10: June-July-August average sea surface temperature in $^{\circ}\text{C}$ difference for Pmin-Pmax (left) and Tmax-Tmin (right). Contours indicate values for Pmax (left) or Tmin (right). Results for all models can be found in the Supplementary Materials (Figure C.14).

357 The increased precipitation, reduced surface pressure and increased convec-
 358 tion over the western Indian Ocean during Pmin are characteristic of a positive
 359 Indian Ocean Dipole (IOD) pattern (Saji et al., 1999). Surface winds along the
 360 equator are more westward (Figure 7), conceivably forced westward by the high
 361 surface pressure anomaly over south-eastern Asia (Figure 5). Because of the
 362 more westward winds, there is more upwelling in the east near Sumatra, and
 363 warm waters reach further west, reducing the east-west sea surface tempera-
 364 ture gradient over the tropical Indian Ocean (Figure 10). Warmer sea surface
 365 temperatures in the western Indian Ocean reduce surface pressure and sup-
 366 port increased convection. Furthermore, cooler sea surface temperatures in the
 367 north-western Arabian Sea, at the coast of Oman, are indicative of more up-
 368 welling due to stronger north-eastward monsoon winds during Pmin. Similar
 369 sea surface temperature changes are produced by GFDL and CESM (Figure
 370 C.14), with particularly strong cooling west of Sumatra in CESM which could
 371 be related to relatively strong east-west sea level pressure difference in CESM
 372 as well as a strong increase in westward winds (however, note that the wind for
 373 CESM is plotted at the lowest pressure level, roughly 66m above the surface
 374 instead of at 10m as for the other models, Figure C.7).

375 **3.2 Obliquity**

376 Obliquity-induced insolation changes are smaller than precession-induced changes,
377 in line with the insolation forcing. The JJA averaged insolation between 10°N
378 and 40°N is $\sim 6 \text{ Wm}^{-2}$ higher during Tmax than Tmin. At the same time SH
379 insolation is reduced, creating an increased interhemispheric insolation gradient
380 (Figure 1) (Bosmans et al., 2015b).

381 Summer precipitation is slightly higher during Tmax over monsoonal Asia,
382 on the order of 0.5 mm/day (Figure 2). Precipitation patterns during Tmax
383 and Tmin are quite similar, but during Tmax precipitation is increased just
384 south of the Tibetan plateau, parts of south-eastern Asia and over the western
385 Indian Ocean. There is inter-model spread in the pattern of change, which can
386 at least partly be explained by differences in the control experiment (Tmin,
387 contours in Figure 3 on the right). For instance, the precipitation maxima
388 over the eastern Indian Ocean is located in different locations, but all models
389 show decreased precipitation during Tmax over these locations. CESM shows
390 decreases in precipitation over eastern China and south-east of the Tibetan
391 plateau, which could be related to CESM's high precipitation rates during Tmin
392 in these areas.

393 Summer temperatures are higher north of 25-30°N during Tmax, because
394 of the small heat capacity of the continent and the fact that the NH insolation
395 increase is stronger towards the higher latitudes. Over India and South-East
396 Asia temperatures are slightly lower because of increased cloud cover, especially
397 over Pakistan and India in the NH (Figure 4). In the SH temperatures are lower
398 due to decreased JJA insolation during Tmax. Some parts of the Indian Ocean
399 do not show a cooling during Tmax in CESM and GFDL (Figure C.1).

400 Changes in surface pressure roughly follow the temperature changes over
401 the continent and the Indian Ocean; surface pressure is lower over the conti-
402 nent north of $\sim 25^\circ\text{N}$ and higher south of $\sim 25^\circ\text{N}$ (Figure 6). As for Pmin, the
403 North Pacific High is stronger during Tmax (Figure 6). Over southern India
404 / the northern Indian Ocean pressure is also slightly increased in EC-Earth

405 and CESM (Figure C.5). Related to the decreased meridional pressure gradient
406 over this area, wind speeds are decreased and slightly more westward (Figure 7).
407 Just south of the equator wind speeds are increased, especially west of Sumatra.
408 Monsoon winds in the northern Arabian Sea, Bay of Bengal and Chinese Seas
409 are increased, in line with stronger monsoons during Tmax. Stronger monsoon
410 winds over East Asia are in agreement with a stronger east-west pressure gradi-
411 ent, as surface pressure is reduced over land and increased over the North Pacific
412 (Figure 6). Similar patterns over the coasts of the monsoon areas emerge from
413 all models (Figure C.7), with some model differences in wind speed changes over
414 the Indian Ocean and the coasts of East Asia.

415 The stronger monsoon winds over the coasts bring more moisture into the
416 continent; moisture transport over these regions is generally increased (Figure
417 6). Over the south-western tropical Indian Ocean moisture transport is slightly
418 reduced, due to both weaker winds and reduced specific humidity (not shown).
419 The latter is related to reduced JJA insolation and lower temperatures over the
420 tropics and the SH during Tmax. This decrease in moisture transport as well
421 as the increase over the coast of the ISM area is also displayed by HadCM3 and
422 CESM, while GFDL shows slightly stronger moisture transport over the south-
423 western tropical Indian Ocean (Figures C.4, C.5, C.6). Over the East Asian
424 coasts, moisture transport into the EASM area is increased in all models, with
425 some inter-model difference in the direction of change. Further model difference
426 occurs in the moisture transport from the tropical Pacific Ocean, which does not
427 occur in EC-Earth and CESM but does occur in GFDL and HadCM3. There
428 does not seem to be a consistent difference in surface pressure changes over the
429 tropical and southern Pacific ocean to accompany these inter-model differences
430 in moisture transport.

431 Changes in evaporation over both land and sea are small (Figure 8). This
432 supports our finding that the increased monsoonal precipitation during Tmax is
433 not related to increased local recycling over land nor to enhanced nearby ocean
434 evaporation, but to a redistribution of precipitation from ocean to land and

435 changes in moisture transport. A small increase in evaporation and moisture
436 transport occurs over the southern tropical Indian Ocean (Figures 8, 6), but not
437 in HadCM3 (Figure C.8). The latter could be related to HadCM3 producing
438 increased precipitation over the western Indian Ocean much further east than
439 the other models (see for instance net precipitation in Figure C.6). The re-
440 duced moisture transport from the tropical Pacific is related to both changes in
441 wind as well as specific humidity in EC-Earth (see Supplementary Figure C.9).
442 The reduced moisture transport over this area in CESM is mostly related to
443 wind (Figure C.11), as is the increased moisture transport displayed by GFDL
444 and HadCM3 (Figures C.10, C.12). In the moisture transport over the coasts
445 into the ISM and EASM area, wind changes play a major role in all models
446 (see Supplementary Figure C.9 and C.10, C.11, C.12), with stronger southerly
447 flow over the EASM and more westward flow over the Indian Ocean related to
448 anomalously high pressure (Figure 6).

449 The vertical velocity at 500 hPa (Figure 9) further shows the redistribution
450 of precipitation: upward velocity (convection) is reduced over the oceans and
451 increased over land, mostly over the regions with the strongest precipitation
452 increase (Figure 3). The exception to this land / ocean response is the west-
453 ern tropical Indian Ocean, where during Tmax convection is slightly stronger
454 and precipitation is higher. This pattern of vertical velocity change, overlaying
455 precipitation changes, can also be seen in all models (Figure C.13). The In-
456 dian Ocean Dipole (IOD)-like pattern is similar to the Pmin-Pmax anomalies
457 described in Section 3.1, with more westward winds along the equator and a
458 reduced east-west sea surface temperature gradient. Sea surface temperatures
459 are overall lower during Tmax due to reduced JJA insolation over most of the
460 tropics and SH in EC-Earth. A colder sea surface is also a reason for the lack of
461 decreased surface pressure over the western Indian Ocean (Figure 6). Nonethe-
462 less the cooling effect of increased upwelling during Tmax can be seen in the east,
463 near Sumatra, as well as over the north-western Arabian Sea, near the coast of
464 Oman where winds are stronger (Figure 10, 7). These upwelling features can

465 be seen in other models as well, despite differences in sea surface temperature
466 change. CESM and GFDL show slightly warmer temperatures over parts of the
467 Indian Ocean (Figure C.14).

468 **4 Discussion**

469 This is the first study to investigate the separate effects of precession and obliquity
470 at high resolution using multiple GCMs. We have shown that monsoon
471 precipitation is enhanced over Asia during minimum precession and maximum
472 obliquity (Pmin and Tmax), when summer insolation in the Northern Hemisphere
473 (NH) is increased. Here we discuss how our results compare to previous
474 modelling studies, how the responses to precession and obliquity differ, and the
475 possible implications for proxy climate studies of the Asian monsoons.

476 **4.1 Previous model studies**

477 Overall, the strengthening of the Asian monsoons at times of precession-induced
478 increased NH summer insolation is recognized in many paleoclimate modelling
479 studies. The mid-Holocene is often used for orbital studies and is a selected
480 timeslice of the Paleoclimate Modelling Intercomparison Project (PMIP), when
481 perihelion occurred in autumn and the insolation difference compared to present-day
482 is similar to but of smaller amplitude than the Pmin-Pmax difference used
483 here. In a Mid-Holocene study performed with EC-Earth, the same model
484 version as used here, we therefore found similar but smaller changes compared to
485 the precession-induced changes reported in this present study (Bosmans et al.,
486 2012). These changes are consistent with other PMIP studies which overall
487 report enhanced southerly monsoon winds over East Asia related to an enhanced
488 land-sea thermal contrast and increased pressure over the Pacific as well as
489 increased convection over land (Jiang et al., 2013; Tian and Jiang, 2013; Wang
490 and Wang, 2013; Zheng et al., 2013; Sun et al., 2015). The increased surface
491 pressure along south-east Asia and the Indian Ocean Dipole (IOD) pattern

492 of increased precipitation over the western Indian Ocean are reported for the
493 Mid-Holocene as well (Zhao et al., 2005; Jiang et al., 2013). There is however
494 some model spread in location and magnitude of changes (Zhao et al., 2005;
495 Wang and Wang, 2013). Abram et al. (2007) find stronger IOD events during
496 the Mid-Holocene in model simulations as well as sea surface temperature and
497 precipitation proxy records.

498 The few studies that also focus on idealized extreme precession forcing report
499 enhanced monsoon precipitation over India and East Asia (Erb et al., 2013;
500 Mantsis et al., 2013), but do not discuss the Asian monsoon in detail. However,
501 Mantsis et al. (2013) as well as Wu et al. (2016) provide an explanation for the
502 strengthened North Pacific High at times of enhanced summer insolation. This
503 strengthening is forced locally through decreased latent heat release over the
504 ocean and a more stable air column, as well as remotely through diabatic heating
505 over monsoon areas where latent heat release is increased, in line with stronger
506 monsoon precipitation. Wang et al. (2012) also identified a strengthened North
507 Pacific High during minimum precession, related to tropospheric cooling which
508 is suggested to be related reduced local latent heat release as well as to land
509 surface heating. Higher surface pressure over the North Pacific is also modeled
510 by Shi et al. (2011). Moreover, Mantsis et al. (2013) display an IOD pattern
511 in their precipitation anomalies over the tropical Indian Ocean (Figure 3), as
512 do time slices with minimum precession in Wang et al. (2012), Battisti et al.
513 (2014) Rachmayani et al. (2016) and Erb et al. (2015), the latter using the same
514 GFDL model output used here. Wang et al. (2012) furthermore show enhanced
515 westward moisture transport from the tropical Pacific, and Battisti et al. (2014)
516 show enhanced westward winds.

517 In idealized experiments of high (maximum) and low (minimum) obliquity
518 using the same GFDL model output used here, Erb et al. (2013) display weak-
519 ened NH monsoons over northern Africa, India, and parts of China during low
520 obliquity. Chen et al. (2011b) also investigate the effect of obliquity on the Asian
521 monsoons, reporting increased summer precipitation over India and south-east

522 Asia during high obliquity. They further suggest a dipole pattern over eastern
523 Asia, with decreased north-east Asian precipitation during high obliquity. Al-
524 though we see a small area of precipitation decrease over north-east Asia during
525 Tmax as well, this “dipole” is not as strong as in Chen et al. (2011b) except
526 for CESM. Furthermore, they do not observe enhanced precipitation over the
527 western Indian Ocean and show a different surface pressure and wind anomaly
528 pattern compared to our obliquity results. These differences may be due to
529 model and / or resolution differences; their study uses a coarse resolution of
530 $\sim 7.5^\circ \times 4^\circ$. The obliquity experiments of Tuentner et al. (2003) do not show an
531 IOD-like pattern either, which may also be related to coarse resolution and / or
532 to model shortcomings (Bosmans et al., 2015a). Rachmayani et al. (2016) show
533 a drier northern EASM as in CESM and Chen et al. (2011b), but show drying
534 over most of India and no increased precipitation over the western Indian Ocean
535 unlike most of the obliquity results shown here.

536 Although our model results are in line with other model experiments for
537 precession-induced monsoon changes, there is a larger inter-model spread in the
538 obliquity-induced monsoon changes, within the models presented here as well
539 as compared to literature. This could at least partly be related to the much
540 weaker insolation forcing associated with obliquity, whereas the large precession-
541 induced forcing results in much more similar responses. The addition of compo-
542 nents that are lacking from our models may result in slightly different responses.
543 Our simulations do not include a dynamic vegetation module. Changing veg-
544 etation patterns can have a small effect on the monsoonal response to orbital
545 forcing (e.g. Dalmeyer et al., 2010; Tian and Jiang, 2013). Furthermore, dy-
546 namic ice sheets are not included and therefore changes in ice sheet volume or
547 area do not play a role in the monsoonal response discussed here. Our findings
548 imply that the ISM and EASM can respond directly to (sub-)tropical insolation
549 changes. A more detailed discussion on how obliquity influences low-latitude cli-
550 mate without a high-latitude influence can be found in Bosmans et al. (2015b).

551 **4.2 Precession vs. obliquity**

552 The precession-induced changes in insolation are different from those induced
553 by obliquity (Tuenter et al., 2003; Bosmans et al., 2015a). During NH summer
554 (JJA), insolation is increased in the northern hemisphere during both Pmin and
555 Tmax, while at the same time in the SH insolation is also increased during Pmin
556 but decreased during Tmax. At first glance the Asian monsoon changes seem
557 very similar, albeit weaker for obliquity. For both a strengthening of the North
558 Pacific High occurs, creating an increased land/sea pressure gradient over East
559 Asia, resulting in stronger northward monsoon winds. There is increased surface
560 pressure over south-eastern Asia, decreased windspeeds over the northern In-
561 dian Ocean and increased precipitation over the tropical western Indian Ocean
562 for both precession and obliquity. Over the southern Pacific Ocean, pressure is
563 increased during Pmin but not during Tmax, which may explain why westward
564 winds and moisture transport are enhanced during Pmin but not during Tmax.
565 There is however disagreement amongst the models in the direction of change in
566 wind and moisture transport from the Pacific. Changes in sea surface tempera-
567 ture are different between precession and obliquity, due to the JJA SH increase
568 in insolation during Pmin and decrease during Tmax. This results in overall
569 warmer sea surface temperatures during Pmin and colder temperatures during
570 Tmax, the latter being the likely cause of the lack of lower surface pressure over
571 the western tropical Indian Ocean during Tmax. We note however that there is
572 some inter-model spread in the obliquity response of Indian Ocean SSTs. Also,
573 lower temperatures result in lower specific humidity and lower moisture trans-
574 port over the the western Indian Ocean, which were increased for Pmin related
575 to higher JJA insolation and temperatures.

576 **4.3 Proxy climate record studies**

577 Our experiments suggest that the ISM and EASM may respond instantaneously
578 to orbital forcing. Comparing our snapshot experiments of orbital extremes

579 directly to transient proxy climate records in terms of phasing is admittedly
580 not straightforward, and we cannot claim that an instantaneous response is
581 always the case since we did not perform transient simulations nor included
582 other boundary conditions such as glacial cycles. However, a direct response of
583 (Asian) monsoons to summer insolation on the orbital time scales is noticed in
584 several studies. Model studies performing transient simulations over multiple
585 orbital cycles find that June-July-August precipitation is in phase with average
586 June insolation (Kutzbach et al., 2008) or June 21st insolation (Weber and
587 Tuenter, 2011). The latter study further shows that for precession the monsoon
588 remains in phase even when ice sheets are included. Recent speleothem oxygen
589 isotope records from South and East Asia (e.g. Wang et al., 2008; Cai et al., 2015;
590 Kathayat et al., 2016), spanning multiple glacial cycles, show no significant lag
591 between the ISM and the EASM and northern hemisphere summer insolation
592 at the precession band. Yet a small offset between models and proxy records
593 remains, with speleothem oxygen isotope records typically in phase with July or
594 July 21st insolation, while model studies suggest that monsoonal precipitation
595 is in phase with June or June 21st insolation,

596 Nevertheless both types of study suggest a much shorter phase lag with re-
597 spect to precession than previously suggested by e.g. Clemens and Prell (2003);
598 Caley et al. (2011) (for an overview see Liu and Shi (2009); Battisti et al.
599 (2014); Wang et al. (2014, 2017)). Lags of up to 9 kyr for precession and 6
600 kyr for obliquity are derived from marine productivity proxies under the as-
601 sumption that productivity is directly related to monsoon wind strength and
602 upwelling. Thus our results suggest that productivity may be related to other
603 processes (see also Ziegler et al. (2010)). Le Mézo et al. (2016) have recently
604 shown that productivity is not necessarily enhanced at times of a stronger ISM
605 during the last glacial-interglacial cycle. Furthermore, we find that not only
606 upwelling over the western Arabian Sea but also evaporation and latent heat
607 release from the southern tropical Indian Ocean can respond instantaneously to
608 increased northern hemisphere insolation. Therefore, we do not agree with the

609 pronounced lag and mechanisms of the ISM in the precession band in the late
610 Pleistocene proposed by e.g. Clemens and Prell (2003) and Caley et al. (2011),
611 who claim that latent heat export from the southern hemisphere into the ISM
612 region is maximized during Pmax, when SH summer insolation is high (Rud-
613 diman, 2006a). According to these mechanisms, the ISM should be stronger
614 during Pmax. The recent speleothem records mentioned above also disagree
615 with this mechanism, with Kathayat et al. (2016) stating that their results do
616 not suggest a dominant influence on the ISM of southern hemisphere climate
617 processes. We do note, however, that the discussion on interpreting cave oxygen
618 isotope records is ongoing (Caley et al., 2014; Mohtadi et al., 2016; Wang et al.,
619 2017).

620 Further investigation into a possible lag in the response time of the Asian
621 monsoons to orbital forcing is necessary. An alternative explanation for the dis-
622 crepancy in model studies which do not find lags and the range of lags found in
623 proxy records is that monsoons may respond more strongly to a phase of preces-
624 sion other than maximum or minimum precession (e.g. Marzin and Braconnot,
625 2009; Erb et al., 2015). For example, if the strongest monsoons are produced
626 when perihelion occurs sometime after the summer solstice, this will appear as
627 a lag with respect to the precession parameter in the proxy record even if the
628 climate system is directly responding to the imposed forcing (see e.g. Figure
629 3 in Erb et al. (2015)). Another aspect that may appear as a lag in the mon-
630 soon strength relative to insolation is the interruption by cold spells such as the
631 Younger Dryas or meltwater spikes in the North Atlantic affecting meridional
632 overturning (e.g. Wang et al., 2008; Ziegler et al., 2010; Mohtadi et al., 2016;
633 Cheng et al., 2016). Such events could cause a longer, up to 3 kyr, lag during
634 major deglaciation. Like ice sheet variations these aspects are not included in
635 this model study. Additional time slice or transient experiments, including ice
636 sheets and potentially Atlantic meltwater fluxes, could shed more light on this
637 discussion.

638 **5 Conclusion**

639 This study set out to investigate the effects of both precession and obliquity on
640 the Asian summer monsoons, using four fully coupled general circulation mod-
641 els; EC-Earth, GFDL, CESM and HadCM3. We demonstrate the effect of both
642 precession and obliquity on the Asian summer monsoons, with increased mon-
643 soon precipitation and convection over the continent during minimum precession
644 and maximum obliquity related to wind-driven changes in moisture transport.
645 Over East Asia the southerly monsoon flow and moisture transport is strength-
646 ened by an intensified North Pacific High and the subsequent increase in the
647 land/sea pressure gradient. Over the Indian monsoon region changes are less
648 straightforward. Anomalously high pressure over south-east Asia weakens the
649 monsoon winds over most of the northern Indian Ocean, reducing evaporation.
650 Over the tropical Indian Ocean an Indian Ocean Dipole pattern emerges with
651 enhanced precipitation over the western Indian Ocean. Therefore these effects
652 damp the enhanced landward moisture transport and monsoonal precipitation
653 over the continent. The influence of obliquity is smaller than that of precession,
654 and shows a different response in temperature and humidity over the Indian
655 Ocean due to reduced insolation over the southern hemisphere. However, for
656 both precession and obliquity wind speed and evaporation is increased over the
657 southern Indian Ocean. For precession, the western tropical Pacific acts as a
658 moisture source as well. Wind speed, and therefore also upwelling, is increased
659 near the coast of Oman. Our results thus show that a direct response to pre-
660 cession and obliquity forcing is possible, in line with speleothem records but in
661 contrast to marine proxy climate records, which suggest a significantly longer
662 lag in response.

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683 **References**

- 684 Abram, N. J., Gagan, M. K., Liu, Z., Hantoro, W. S., McCulloch, M. T., and
685 Suwargadi, B. W.: Seasonal characteristics of the Indian Ocean Dipole during
686 the Holocene epoch, *Nature*, 445, 299–302, 2007.
- 687 Araya-Melo, P. A., Crucifix, M., and Bounceur, N.: Global sensitivity analysis
688 of the Indian monsoon during the Pleistocene, *Climate of the Past*, 11, 45–61,
689 2015.
- 690 Battisti, D., Ding, Q., and Roe, G.: Coherent pan-Asian climatic and isotopic
691 response to orbital forcing of tropical insolation, *Journal of Geophysical Re-*
692 *search: Atmospheres*, 119, 2014.

- 693 Berger, A. L.: Long-Term Variations of Daily Insolation and Quaternary Cli-
694 matic Changes, *Journal of the Atmospheric Sciences*, 35, 2362–2367, 1978.
- 695 Bordoni, S. and Schneider, T.: Monsoons as eddy-mediated regime transitions
696 of the tropical overturning circulation, *Nature Geoscience*, 1, 515, 2008.
- 697 Bosmans, J., Drijfhout, S., Tuenter, E., Hilgen, F., and Lourens, L.: Response
698 of the North African summer monsoon to precession and obliquity forcings in
699 the EC-Earth GCM, *Climate Dynamics*, 44, 279–297, 2015a.
- 700 Bosmans, J., Hilgen, F., Tuenter, E., and Lourens, L.: Obliquity forcing of
701 low-latitude climate, *Climate of the Past*, 11, 1335–1346, 2015b.
- 702 Bosmans, J. H. C., Drijfhout, S. S., Tuenter, E., Lourens, L. J., Hilgen,
703 F. J., and Weber, S. L.: Monsoonal response to mid-holocene orbital forc-
704 ing in a high resolution GCM, *Climate Of The Past*, 8, 723–740, doi:
705 10.5194/cp-8-723-2012, 2012.
- 706 Braconnot, P. and Marti, O.: Impact of precession on monsoon charac-
707 teristics from coupled ocean atmosphere experiments: changes in Indian
708 monsoon and Indian ocean climatology, *Marine Geology*, 201, 23–34, doi:
709 10.1016/S0025-3227(03)00206-8, 2003.
- 710 Braconnot, P., Otto-Bliesner, B., Harrison, S., Joussaume, S., Peterchmitt,
711 J. Y., Abe-ouchi, A., Crucifix, M., Driesschaert, E., Fichefet, T., Hewitt,
712 C. D., Kageyama, M., Kitoh, A., Laine, A., Loutre, M. F., Marti, O., Merkel,
713 U., Ramstein, G., Valdes, P., Weber, S. L., Yu, Y., and Zhao, Y.: Results
714 of PMIP2 coupled simulations of the Mid-Holocene and Last Glacial Maxi-
715 mum – Part 1: experiments and large-scale features, *Climate Of The Past*, 3,
716 261–277, 2007.
- 717 Braconnot, P., Marzin, C., Gregoire, L., Mosquet, E., and Marti, O.: Monsoon
718 response to changes in Earth’s orbital parameters: comparisons between sim-
719 ulations of the Eemian and of the Holocene, *Climate Of The Past*, 4, 281–294,
720 2008.

- 721 Cai, Y., Fung, I. Y., Edwards, R. L., An, Z., Cheng, H., Lee, J.-E., Tan, L.,
722 Shen, C.-C., Wang, X., Day, J. A., et al.: Variability of stalagmite-inferred
723 Indian monsoon precipitation over the past 252,000 y, *Proceedings of the*
724 *National Academy of Sciences*, 112, 2954–2959, 2015.
- 725 Caley, T., Malaizé, B., Zaragosi, S., Rossignol, L., Bourget, J., Eynaud,
726 F., Martinez, P., Giraudeau, J., Charlier, K., and Ellouz-Zimmermann,
727 N.: New Arabian Sea records help decipher orbital timing of Indo-Asian
728 monsoon, *Earth and Planetary Science Letters*, 308, 433 – 444, doi:<http://dx.doi.org/10.1016/j.epsl.2011.06.019>, URL <http://www.sciencedirect.com/science/article/pii/S0012821X11003785>, 2011.
- 731 Caley, T., Roche, D. M., and Renssen, H.: Orbital Asian summer monsoon
732 dynamics revealed using an isotope-enabled global climate model, *Nature*
733 *communications*, 5, 2014.
- 734 Cattle, H., Crossley, J., and Drewry, D.: Modelling arctic climate change, *Philosophical Transactions of the Royal Society of London A: Mathematical, Physical and Engineering Sciences*, 352, 201–213, 1995.
- 737 Chen, G.-S., Kutzbach, J., Gallimore, R., and Liu, Z.: Calendar effect on phase
738 study in paleoclimate transient simulation with orbital forcing, *Climate dynamics*, 37, 1949–1960, 2011a.
- 740 Chen, G.-s., Zhengyu, L., Clemens, S. C., Prell, W. L., and Liu, X.: Modeling the time-dependent response of the Asian summer monsoon to obliquity
741 forcing in a coupled GCM: a PHASEMAP sensitivity experiment, *Climate*
742 *Dynamics*, 36, 695–710, doi:10.1007/s00382-010-0740-3, 2011b.
- 744 Cheng, H., Edwards, R. L., Sinha, A., Spötl, C., Yi, L., Chen, S., Kelly, M.,
745 Kathayat, G., Wang, X., Li, X., et al.: The Asian monsoon over the past
746 640,000 years and ice age terminations, *Nature*, 534, 640–646, 2016.
- 747 Clemens, S. C. and Prell, W. L.: A 350,000 year summer-monsoon multi-proxy

- 748 stack from the Owen Ridge, Northern Arabian Sea, *Marine Geology*, 201,
749 35–51, 2003.
- 750 Dallmeyer, A., Claussen, M., and Otto, J.: Contribution of oceanic and vege-
751 tation feedbacks to Holocene climate change in monsoonal Asia, *Climate Of*
752 *The Past*, 6, 195–218, 2010.
- 753 Delworth, T. L., Broccoli, A. J., Rosati, A., Stouffer, R. J., Balaji, V., Beesley,
754 J. A., Cooke, W. F., Dixon, K. W., Dunne, J., Dunne, K., et al.: GFDL’s
755 CM2 global coupled climate models. Part I: Formulation and simulation char-
756 acteristics, *Journal of Climate*, 19, 643–674, 2006.
- 757 Dolan, A. M., Haywood, A. M., Hill, D. J., Dowsett, H. J., Hunter, S. J., Lunt,
758 D. J., and Pickering, S. J.: Sensitivity of Pliocene ice sheets to orbital forcing,
759 *Palaeogeography, Palaeoclimatology, Palaeoecology*, 309, 98–110, 2011.
- 760 Erb, M. P., Broccoli, A. J., and Clement, A. C.: The contribution of radiative
761 feedbacks to orbitally-driven climate change, *Journal of Climate*, doi:doi:10.
762 1175/JCLI-D-12-00419.1, 2013.
- 763 Erb, M. P., Jackson, C. S., and Broccoli, A. J.: Using single-forcing GCM
764 simulations to reconstruct and interpret Quaternary climate change, *Journal*
765 *of Climate*, 28, 9746–9767, 2015.
- 766 Erb, M. P., Jackson, S., Broccoli, A. J., Lea, D. W., Valdes, P. J., Crucifix, M.,
767 and DiNezio, P. M.: Model evidence for a seasonal bias in Antarctic ice cores,
768 in press, revisions submitted, 2018.
- 769 Gordon, C., Cooper, C., Senior, C. A., Banks, H., Gregory, J. M., Johns, T. C.,
770 Mitchell, J. F., and Wood, R. A.: The simulation of SST, sea ice extents
771 and ocean heat transports in a version of the Hadley Centre coupled model
772 without flux adjustments, *Climate Dynamics*, 16, 147–168, 2000.
- 773 Hazeleger, W., Severijns, C., Semmler, T., Stefanescu, S., Yang, S., Wyser,
774 K., Wang, X., Dutra, E., Baldasano, J. M., Bintanja, R., Bougeault, P.,

775 Caballero, R., Ekman, A. M., Christensen, J. H., van den Hurk, B., Jimenez,
776 P., Jones, C., Kallberg, P., Koenigk, T., McGrath, R., Miranda, P., van Noije,
777 T., Palmer, T., Parodi, J. A., Schmith, T., Selten, F., Storelmo, T., Sterl,
778 A., Tapamo, H., Vancoppenolle, M., Viterbo, P., and Willen, U.: EC-Earth:
779 A Seamless Earth System Prediction Approach in Action, *Bulletin of the*
780 *American Meteorological Society*, 91, 1357–1363, 2010.

781 Hazeleger, W., Wang, X., Severijns, C., Stefanescu, S., Bintanja, R., Sterl, A.,
782 Wyser, K., Semmler, T., Yang, S., van den Hurk, B., van Noije, T., van der
783 Linden, E., and van der Wiel, K.: EC-Earth V2.2: description and validation
784 of a new seamless earth system prediction model, *Climate Dynamics*, doi:
785 10.1007/s00382-011-1228-5, 2011.

786 Jiang, D., Tian, Z., and Lang, X.: Mid-Holocene net precipitation changes over
787 China: model–data comparison, *Quaternary Science Reviews*, 82, 104–120,
788 2013.

789 Kathayat, G., Cheng, H., Sinha, A., Spötl, C., Edwards, R. L., Zhang, H., Li,
790 X., Yi, L., Ning, Y., Cai, Y., et al.: Indian monsoon variability on millennial-
791 orbital timescales, *Scientific reports*, 6, 2016.

792 Kutzbach, J. E. and Guetter, P. J.: The Influence of Changing Orbital Param-
793 eters and Surface Boundary Conditions on Climate Simulations for the Past
794 18000 years, *Journal of the Atmospheric Sciences*, 43, 1726–1759, 1986.

795 Kutzbach, J. E. and Otto-Bliesner, B. L.: The Sensitivity of the African-Asian
796 Monsoonal Climate to Orbital Parameter Changes for 9000 Years B.P. in
797 a Low-Resolution General Circulation Model, *Journal of the Atmospheric*
798 *Sciences*, 39, 1177–1188, 1982.

799 Kutzbach, J. E., Liu, X., Liu, Z., and Chen, G.: Simulation of the evolutionary
800 response of global summer monsoons to orbital forcing over the past 280,000
801 years, *Climate Dynamics*, 30, 567–579, doi:10.1007/s00382-007-0308-z, 2008.

- 802 Le Mézo, P., Beaufort, L., Bopp, L., Braconnot, P., and Kageyama, M.: From
803 Monsoon to marine productivity in the Arabian Sea: insights from glacial and
804 interglacial climates, *Climate of the Past Discussion*, doi:10.5194/cp-2016-88,
805 2016.
- 806 Liu, X. and Shi, Z.: Effect of precession on the Asian summer monsoon evolu-
807 tion: A systematic review, *Chinese Science Bulletin*, 54, 3720–3730, 2009.
- 808 Mantsis, D. F., Clement, B., Kirtman, B., Broccoli, A. J., and Erb, M. P.:
809 Precessional cycles and their influence on the North Pacific and North Atlantic
810 summer anticyclones, *Journal of Climate*, doi:10.1175/JCLI-D-12-00343.1,
811 2013.
- 812 Marzin, C. and Braconnot, P.: Variations of Indian and African monsoons in-
813 duced by insolation changes at 6 and 9.5 kyr BP, *Climate Dynamics*, 33,
814 215–231, doi:10.1007/s00382-009-0538-3, 2009.
- 815 Marzocchi, A., Lunt, D., Flecker, R., Bradshaw, C., Farnsworth, A., and Hilgen,
816 F.: Orbital control on late Miocene climate and the North African monsoon:
817 insight from an ensemble of sub-precessional simulations, *Climate of the Past*,
818 11, 1271–1295, 2015.
- 819 Mohtadi, M., Prange, M., and Steinke, S.: Palaeoclimatic insights into forcing
820 and response of monsoon rainfall, *Nature*, 533, 191–199, 2016.
- 821 Molnar, P., Boos, W. R., and Battisti, D. S.: Orographic controls on climate and
822 paleoclimate of Asia: thermal and mechanical roles for the Tibetan Plateau,
823 *Annual Review of Earth and Planetary Sciences*, 38, 2010.
- 824 Pollard, D. and Reusch, D. B.: A calendar conversion method for monthly
825 mean paleoclimate model output with orbital forcing, *Journal of Geophysical*
826 *Research: Atmospheres*, 107, 2002.
- 827 Prell, W. L. and Kutzbach, J. E.: Monsoon Variability Over the Past 150,000
828 Years, *Journal of Geophysical Research*, 92, 8411–8425, 1987.

- 829 Prescott, C. L., Haywood, A. M., Dolan, A. M., Hunter, S. J., Pope, J. O.,
830 and Pickering, S. J.: Assessing orbitally-forced interglacial climate variability
831 during the mid-Pliocene Warm Period, *Earth and Planetary Science Letters*,
832 400, 261–271, 2014.
- 833 Rachmayani, R., Prange, M., and Schulz, M.: Intra-interglacial climate vari-
834 ability: model simulations of Marine Isotope Stages 1, 5, 11, 13, and 15,
835 *Climate of the Past*, 12, 677–695, doi:10.5194/cp-12-677-2016, URL <http://www.clim-past.net/12/677/2016/>, 2016.
- 837 Ruddiman, W. F.: What is the timing of orbital-scale monsoon changes?, *Qua-*
838 *ternary Science Reviews*, 25, 657–658, 2006a.
- 839 Ruddiman, W. F.: Orbital changes and climate, *Quaternary Science Reviews*,
840 25, 3092–3112, 2006b.
- 841 Saji, N., Goswami, B. N., Vinayachandran, P., and Yamagata, T.: A dipole
842 mode in the tropical Indian Ocean, *Nature*, 401, 360–363, 1999.
- 843 Shi, Z., Liu, X., Sun, Y., An, Z., Liu, Z., and Kutzbach, J.: Distinct responses
844 of East Asian summer and winter monsoons to astronomical forcing, *Climate*
845 *of the Past*, 7, 1363–1370, 2011.
- 846 Singarayer, J. S. and Valdes, P. J.: High-latitude climate sensitivity to ice-sheet
847 forcing over the last 120kyr, *Quaternary Science Reviews*, 29, 43–55, 2010.
- 848 Sun, Y., Kutzbach, J., An, Z., Clemens, S., Liu, Z., Liu, W., Liu, X., Shi,
849 Z., Zheng, W., Liang, L., et al.: Astronomical and glacial forcing of East
850 Asian summer monsoon variability, *Quaternary Science Reviews*, 115, 132–
851 142, 2015.
- 852 Tian, Z. and Jiang, D.: Mid-Holocene ocean and vegetation feedbacks over
853 East Asia, *Climate of the Past*, 9, 2153–2171, doi:10.5194/cp-9-2153-2013,
854 URL <http://www.clim-past.net/9/2153/2013/>, 2013.

- 855 Tuentner, E., Weber, S. L., Hilgen, F. J., and Lourens, L. J.: The response of
856 the African summer monsoon to remote and local forcing due to precession
857 and obliquity, *Global and Planetary Change*, 36, 219 – 235, doi:10.1016/
858 S0921-8181(02)00196-0, 2003.
- 859 Valcke, S. and Morel, T.: OASIS3 user guide, Tech. rep.,
860 CERFACS, prism Technical Report, 68pp, available online at
861 http://www.prism.enes.org/Publications/Reports/oasis3_UserGuide_T3.pdf,
862 2006.
- 863 Wang, P., Wang, B., Cheng, H., Fasullo, J., Guo, Z., Kiefer, T., and Liu,
864 Z.: The Global Monsoon across Time Scales: coherent variability of regional
865 monsoons, *Climate of the Past*, 10, 1–46, 2014.
- 866 Wang, P. X., Wang, B., Cheng, H., Fasullo, J., Guo, Z., Kiefer, T., and Liu, Z.:
867 The global monsoon across time scales: Mechanisms and outstanding issues,
868 *Earth-Science Reviews*, 174, 84–121, 2017.
- 869 Wang, T. and Wang, H.: Mid-Holocene Asian summer climate and its responses
870 to cold ocean surface simulated in the PMIP2 OAGCMs experiments, *Journal*
871 *of Geophysical Research: Atmospheres*, pp. 1–12, 2013.
- 872 Wang, Y., Cheng, H., Edwards, R. L., Kong, X., Shao, X., Chen, S., Wu, J.,
873 Jiang, X., Wang, X., and An, Z.: Millennial-and orbital-scale changes in the
874 East Asian monsoon over the past 224,000 years, *Nature*, 451, 1090–1093,
875 2008.
- 876 Wang, Y., Jian, Z., and Zhao, P.: Extratropical modulation on Asian summer
877 monsoon at precessional bands, *Geophysical Research Letters*, 39, 2012.
- 878 Weber, S. and Tuentner, E.: The impact of varying ice sheets and greenhouse
879 gases on the intensity and timing of boreal summer monsoons, *Quaternary*
880 *Science Reviews*, 30, 469–479, 2011.

- 881 Wu, C.-H., Lee, S.-Y., Chiang, J. C., and Hsu, H.-H.: The influence of obliquity
882 in the early Holocene Asian summer monsoon, *Geophysical Research Letters*,
883 43, 4524–4530, 2016.
- 884 Zhao, Y., Braconnot, P., Marti, O., Harrison, S. P., Hewitt, C., Kitoh,
885 A., Liu, A., Mikolajewicz, U., Otto-Bliesner, B., and Weber, S. L.: A
886 multi-model analysis of the role of the ocean on the African and Indian
887 monsoon during the mid-Holocene, *Climate Dynamics*, 25, 777–800, doi:
888 10.1007/s00382-005-0075-7, 2005.
- 889 Zheng, W., Wu, B., He, J., and Yu, Y.: The East Asian Summer Monsoon at
890 mid-Holocene: results from PMIP3 simulations, *Climate of the Past*, 9, 453–
891 466, doi:10.5194/cp-9-453-2013, URL [http://www.clim-past.net/9/453/
892 2013/](http://www.clim-past.net/9/453/2013/), 2013.
- 893 Ziegler, M., Lourens, L. J., Tüenter, E., Hilgen, F., Reichert, G.-J., and We-
894 ber, N.: Precession phasing offset between Indian summer monsoon and
895 Arabian Sea productivity linked to changes in Atlantic overturning cir-
896 culation, *Paleoceanography*, 25, n/a–n/a, doi:10.1029/2009PA001884, URL
897 <http://dx.doi.org/10.1029/2009PA001884>, 2010.

A Supplementary material: Choice of calendar

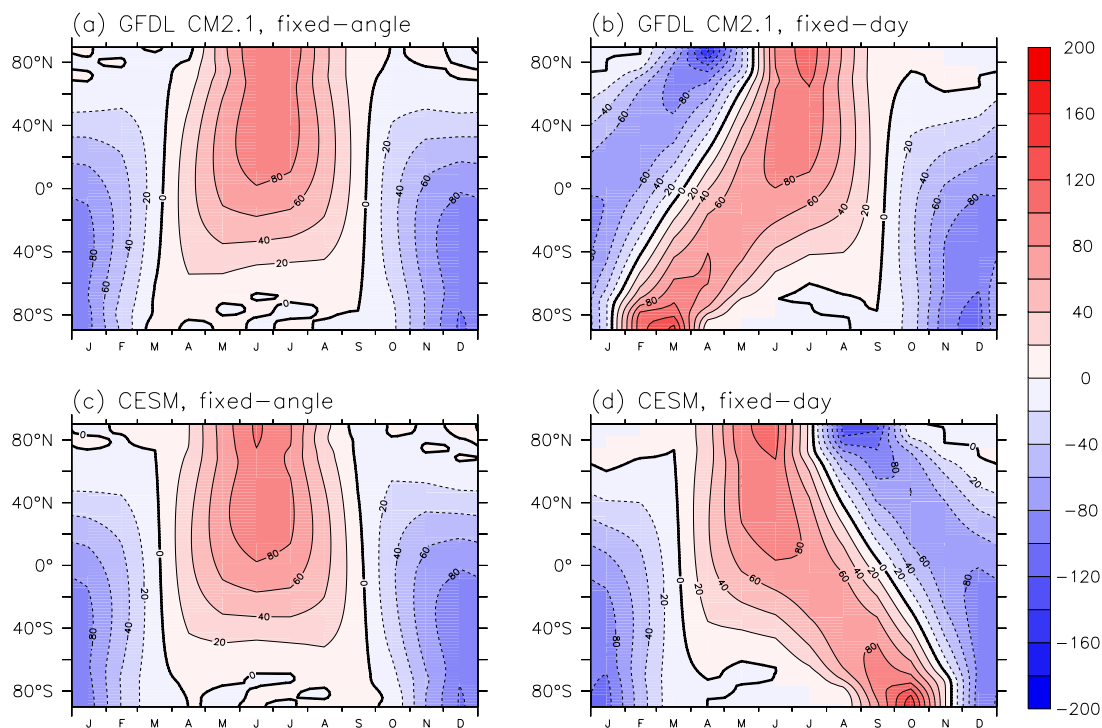


Figure A.1: Precession-induced insolation difference in W/m^2 (Pmin - Pmax) for GFDL (a,b) and CESM (c,d). The fixed-angle calendar used here is shown in (a,c), the original fixed-day calendar in (b,d). Results are converted to the fixed-angle calendar to align solstices and equinoxes throughout the year, but the choice of calendar does not change the conclusions discussed in this paper. Results in (b) and (d) look different primarily because the two models fix the calendar at different dates: the autumnal equinox for GFDL CM2.1 and the vernal equinox for CESM.

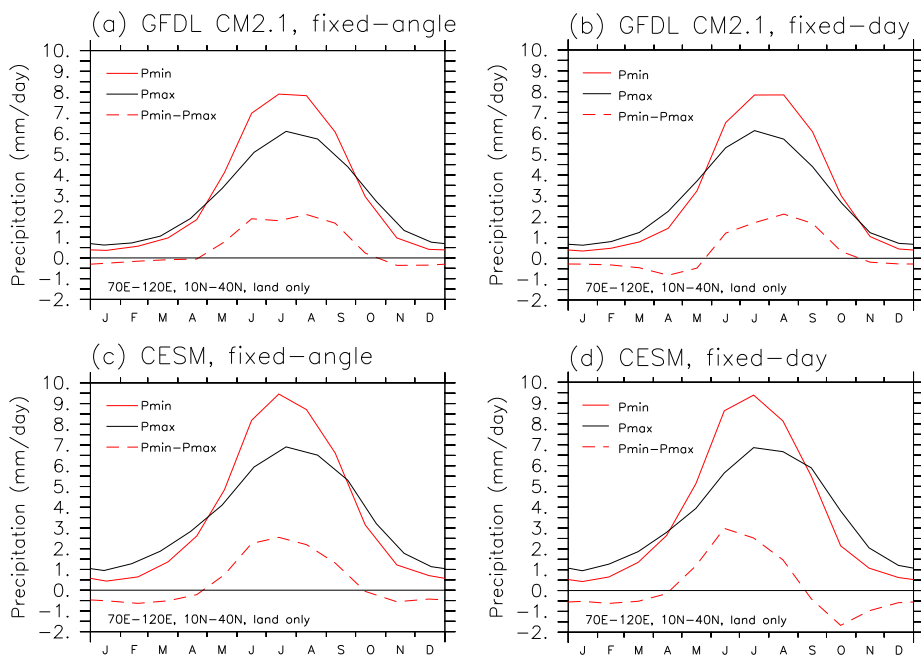


Figure A.2: Precipitation per month for GFDL (a,b) and CESM (c,d) precession experiments, on both the fixed-angle calendar (a,c) and the original fixed-day calendar (b,d). Precipitation is given in mm/day averaged over the area 70°E-120°E, 10°N:40°N, using land grid cells only.

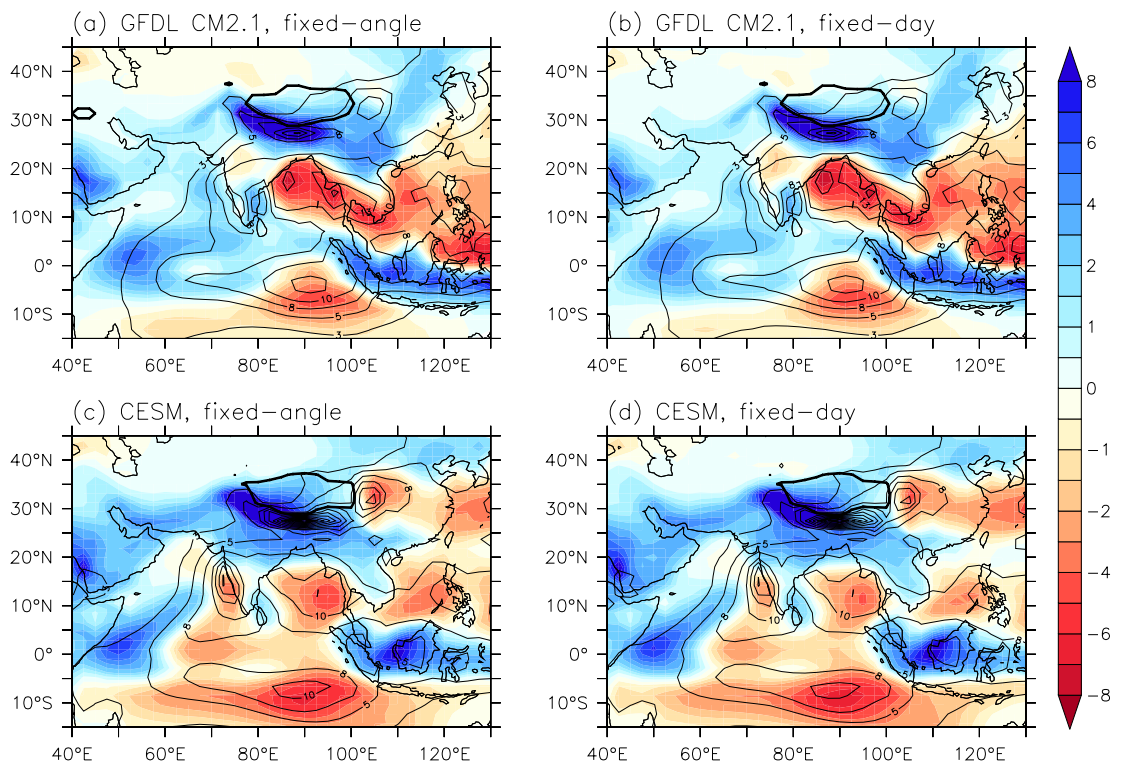


Figure A.3: June-July-August average precipitation difference for Pmin-Pmax for GFDL (a,b) and CESM (c,d). Contours indicate values for Pmax (left) in mm/day. The thick contour line is at 4km height, indicating the Tibetan Plateau. Panels (a,c) show the results on a fixed-angle calendar, and panels (b,d) show the results on the (original) fixed-day calendar.

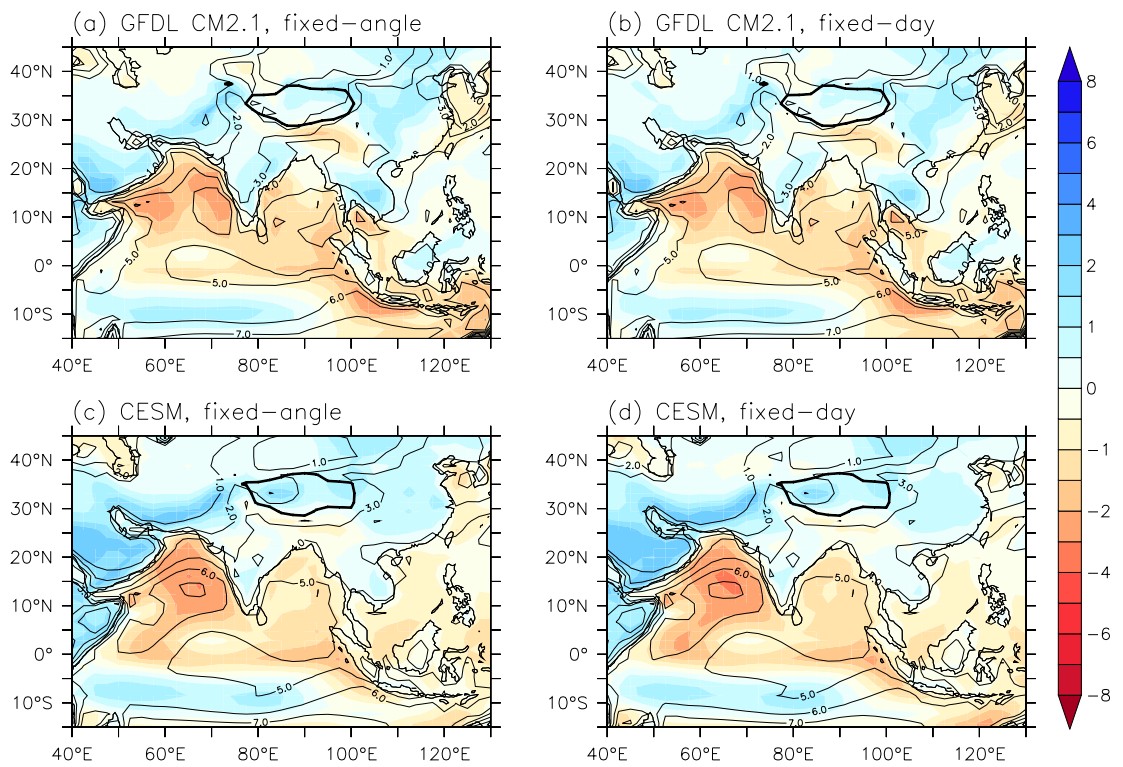


Figure A.4: June-July-August average evaporation difference for Pmin-Pmax for GFDL (a,b) and CESM (c,d). Contours indicate values for Pmax (left) in mm/day. The thick contour line is at 4km height, indicating the Tibetan Plateau. Panels (a,c) show the results on a fixed-angle calendar, and panels (b,d) show the results on the (original) fixed-day calendar.

899 **B Supplementary material: Orography**

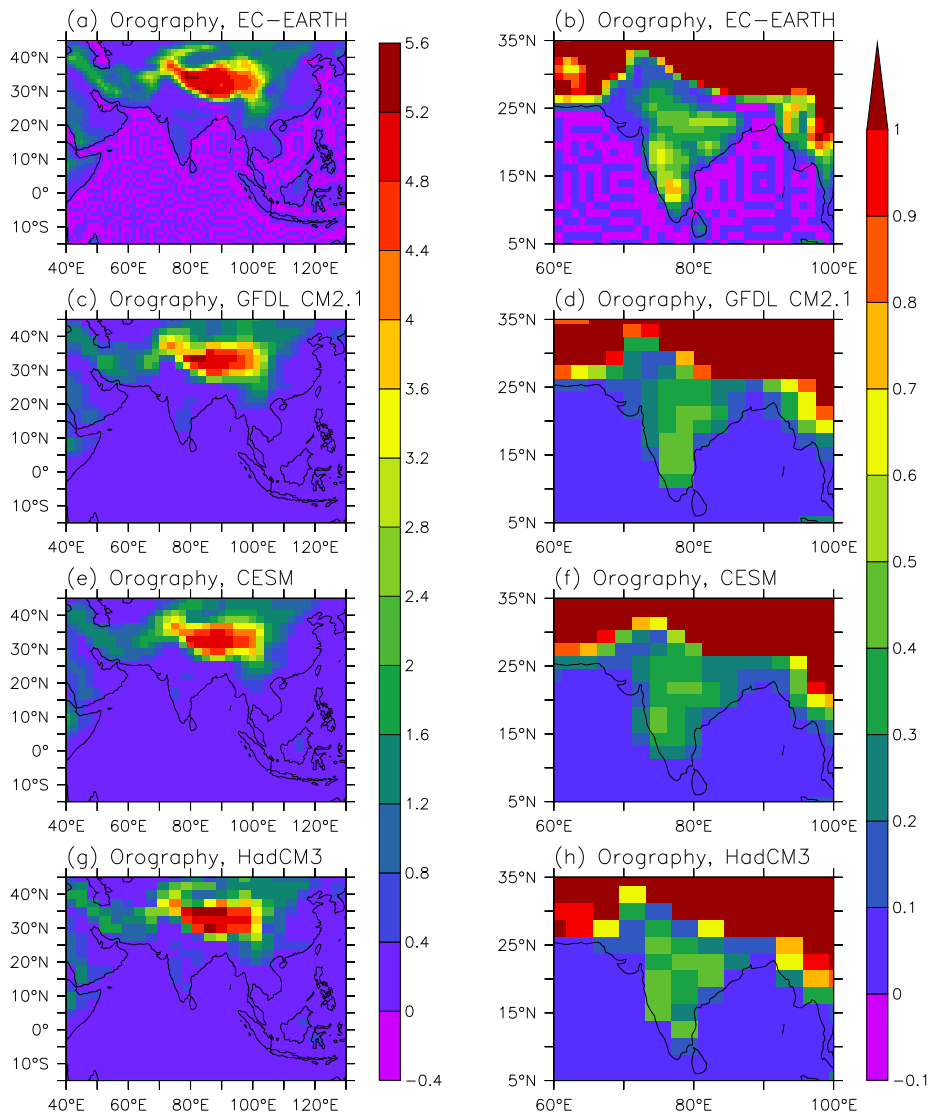


Figure B.1: Surface height in km in all models (orography) over the whole Asian area considered in this study (left) and over India (right). Note the different range in the colour bar left and right.

⁹⁰⁰ **C** **Supplementary material: Results per model**

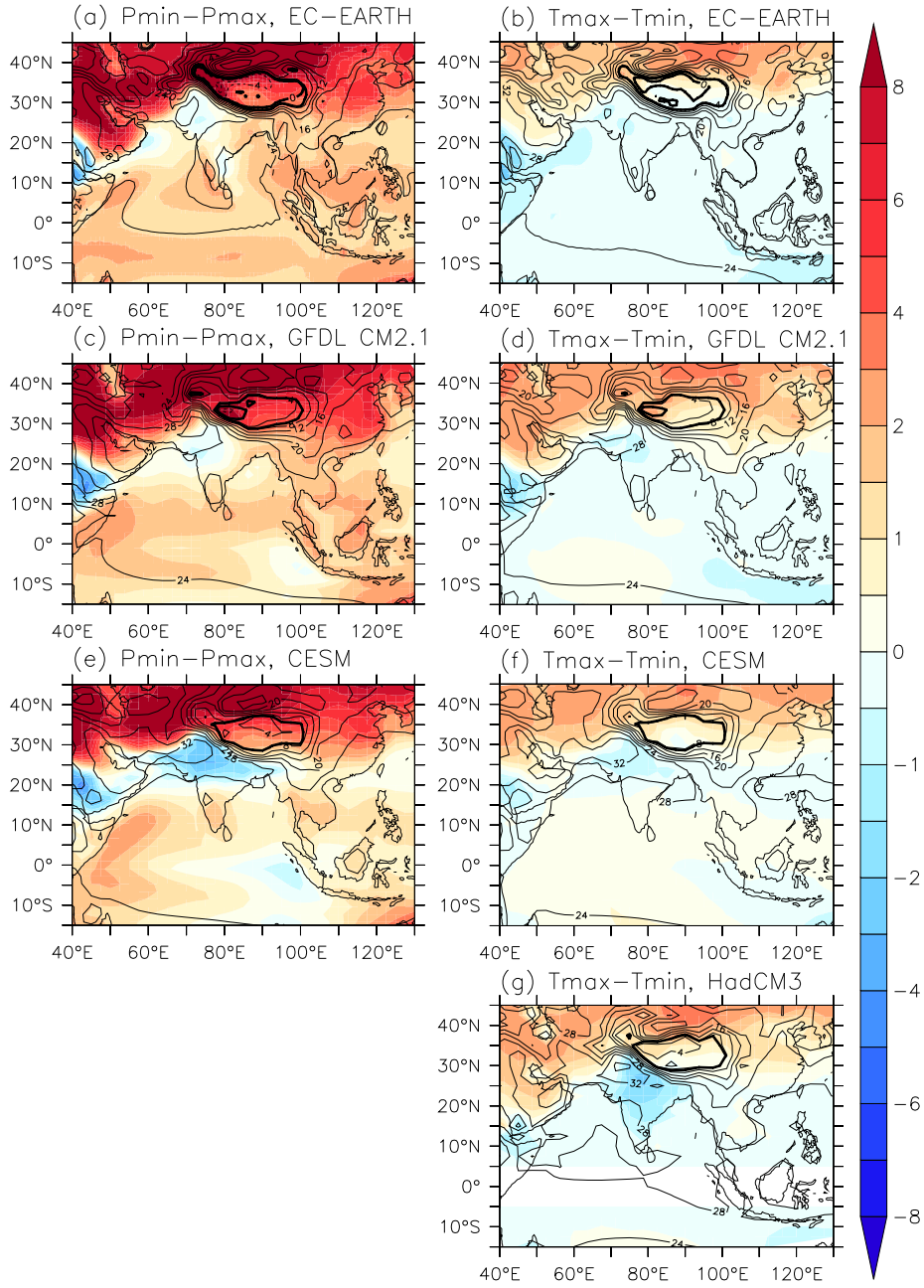


Figure C.1: June-July-August average surface air temperature difference for Pmin-Pmax (left) and Tmax-Tmin (right) for all models. Contours indicate values for Pmax (left) or Tmin (right) in °C. The thick contour line is at 4km height, indicating the Tibetan Plateau. Note that temperature is given at 2m above the surface, except for 1.5m in HadCM3. As in Figure 4 in the main text, but for all models.

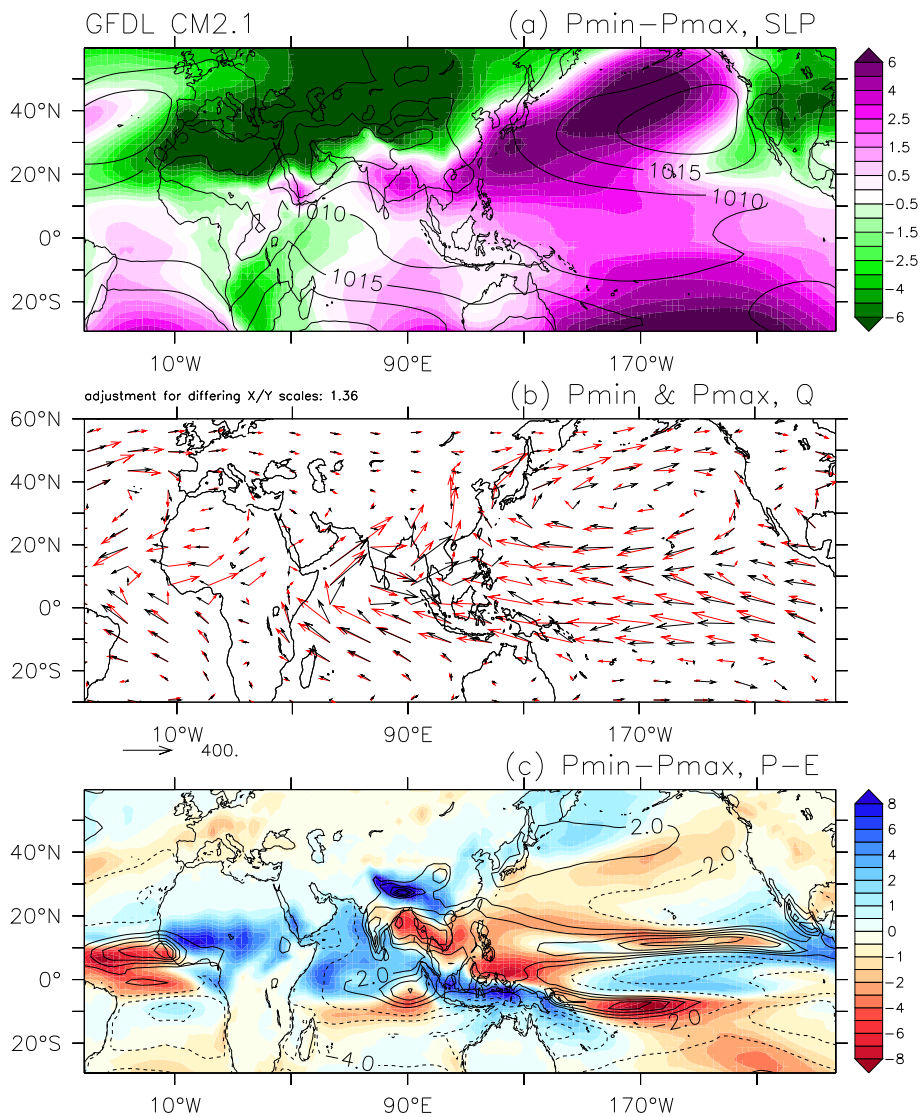


Figure C.2: June-July-August average results for GFDL Pmin-Pmax. Top (a) shows sea level pressure difference in hPa with Pmax values in contours. Middle (b) shows moisture transport Q , the vertical integral of qv in $\text{kg}/(\text{ms})$, during Pmin in red, and Pmax in black. Monthly model outputs are used. Bottom (c) shows net precipitation with positive values (blue) indicating increased net precipitation in mm/day and contours showing net precipitation during Pmax JJA.

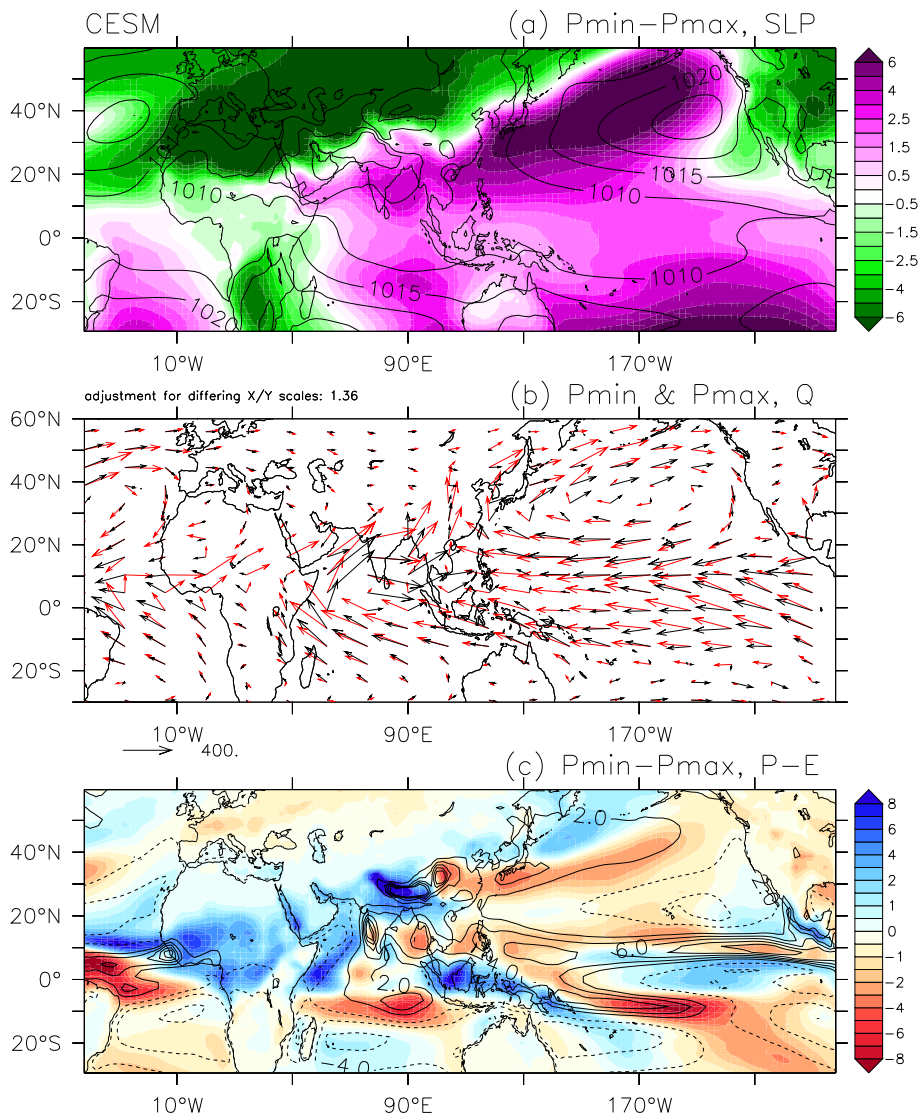


Figure C.3: June-July-August average results for CESM Pmin-Pmax. Top (a) shows sea level pressure difference in hPa with Pmax values in contours. Middle (b) shows moisture transport Q , the vertical integral of qv in $kg/(ms)$, during Pmin in red, and Pmax in black. Monthly model outputs are used. Bottom (c) shows net precipitation with positive values (blue) indicating increased net precipitation in mm/day and contours showing net precipitation during Pmax JJA.

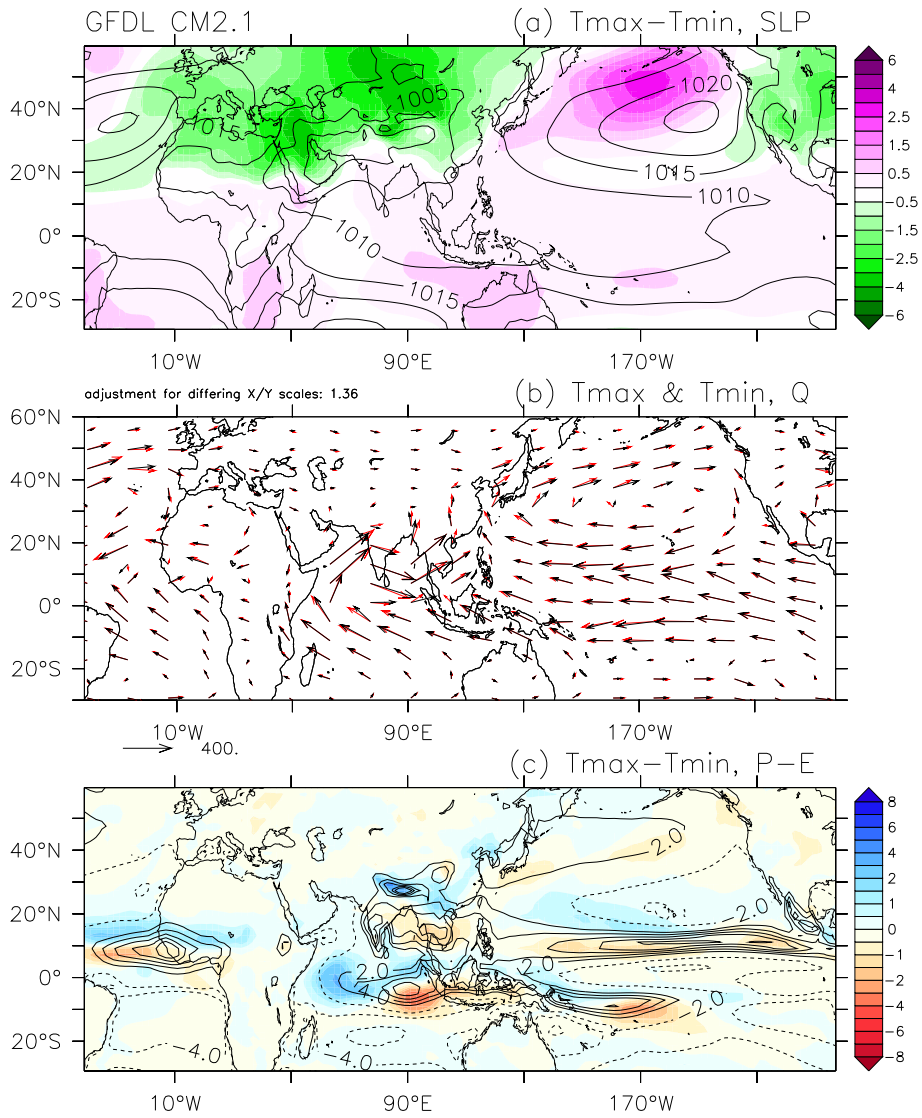


Figure C.4: June-July-August average results for GFDL Tmax-Tmin. Top (a) shows sea level pressure difference in hPa with Tmax values in contours. Middle (b) shows moisture transport Q , the vertical integral of qv in kg/(ms), during Tmax in red, and Tmin in black. Monthly model outputs are used. Bottom (c) shows net precipitation with positive values (blue) indicating increased net precipitation in mm/day and contours showing net precipitation during Tmax JJA.

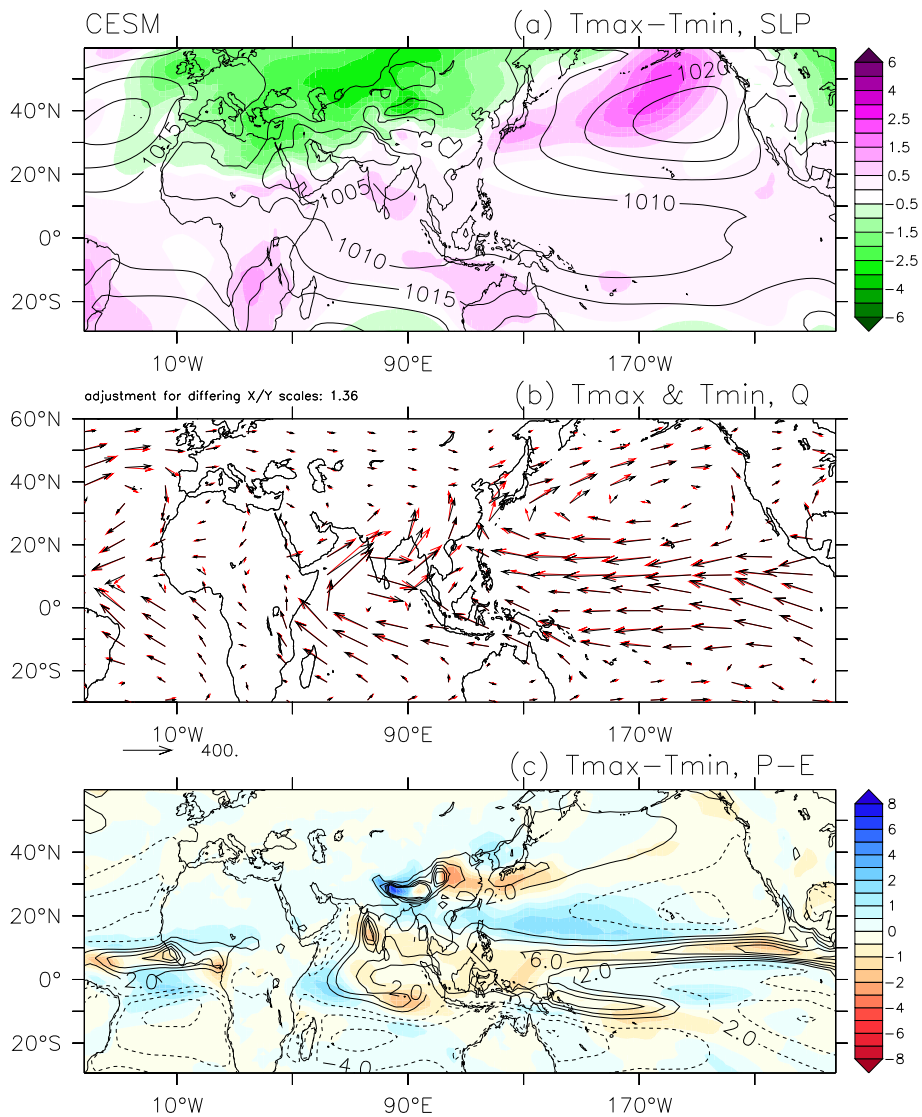


Figure C.5: June-July-August average results for CESM Tmax-Tmin. Top (a) shows sea level pressure difference in hPa with T_{min} values in contours. Middle (b) shows moisture transport Q , the vertical integral of qv in kg/(ms), during T_{max} in red, and T_{min} in black. Monthly model outputs are used. Bottom (c) shows net precipitation with positive values (blue) indicating increased net precipitation in mm/day and contours showing net precipitation during T_{min} JJA.

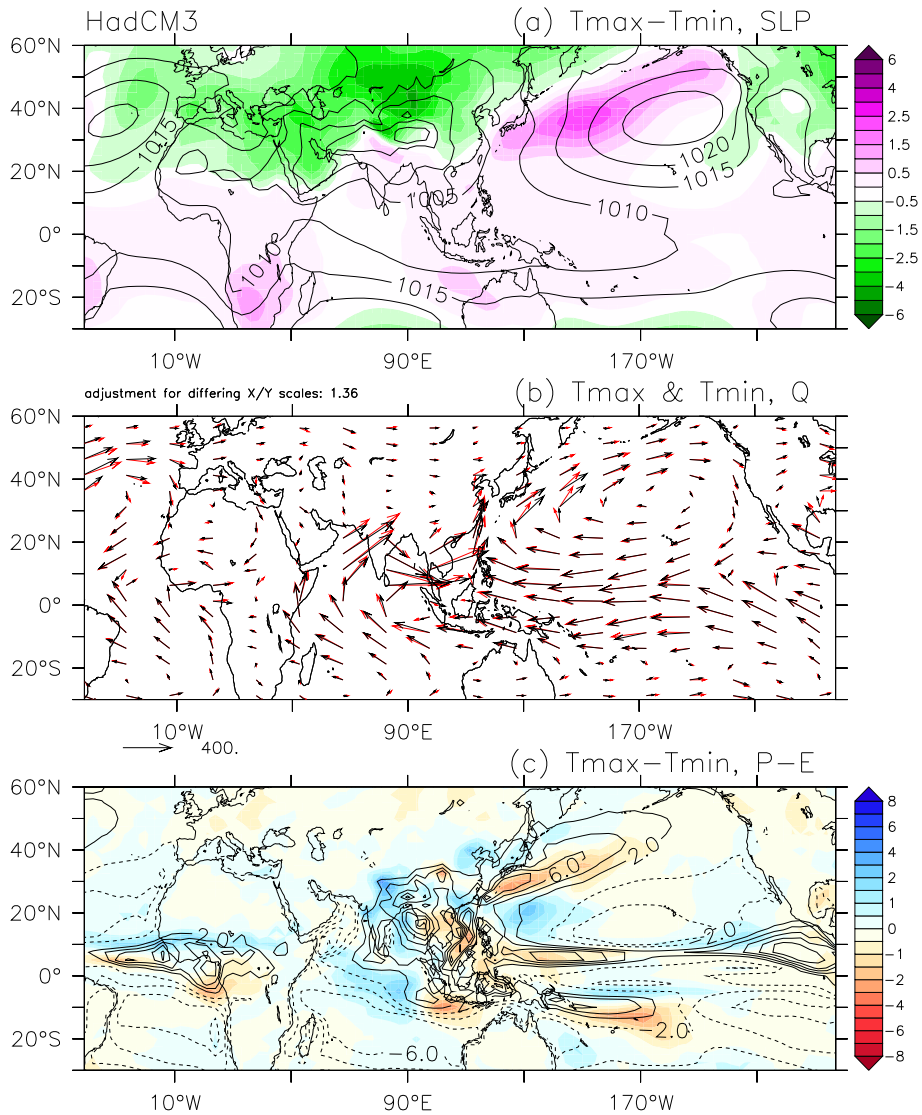


Figure C.6: June-July-August average results for HadCM3 $T_{max}-T_{min}$. Top (a) shows sea level pressure difference in hPa with T_{min} values in contours. Middle (b) shows moisture transport Q , the vertical integral of $q\mathbf{v}$ in $\text{kg}/(\text{ms})$, during T_{max} in red, and T_{min} in black. Monthly model outputs are used. Bottom (c) shows net precipitation with positive values (blue) indicating increased net precipitation in mm/day and contours showing net precipitation during T_{min} JJA.

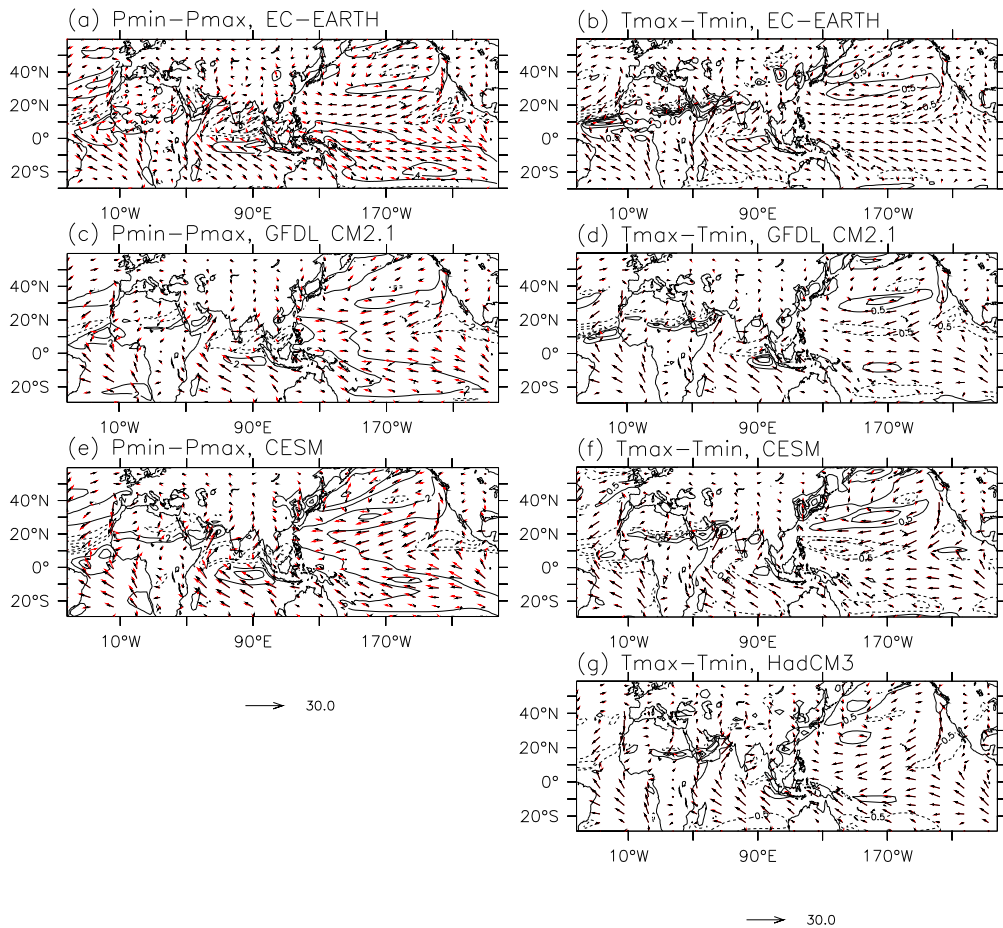


Figure C.7: June-July-August average surface wind in m/s for Pmin-Pmax (left) and Tmax-Tmin (right) for all models. Pmin is given in red, Pmax in black, Tmax in green and Tmin in blue. Contours indicate wind speed differences for Pmin-Pmax (left) and Tmax-Tmin (right), with contour levels set to 2 m/s on the left and 0.5 m/s on the right. Unit length is 30 m/s. Note that wind speed is given at 10m above the surface, except for CESM where only the lowest model level was available, on average 66m above the surface. As in Figure 7 in the main text, but for all models.

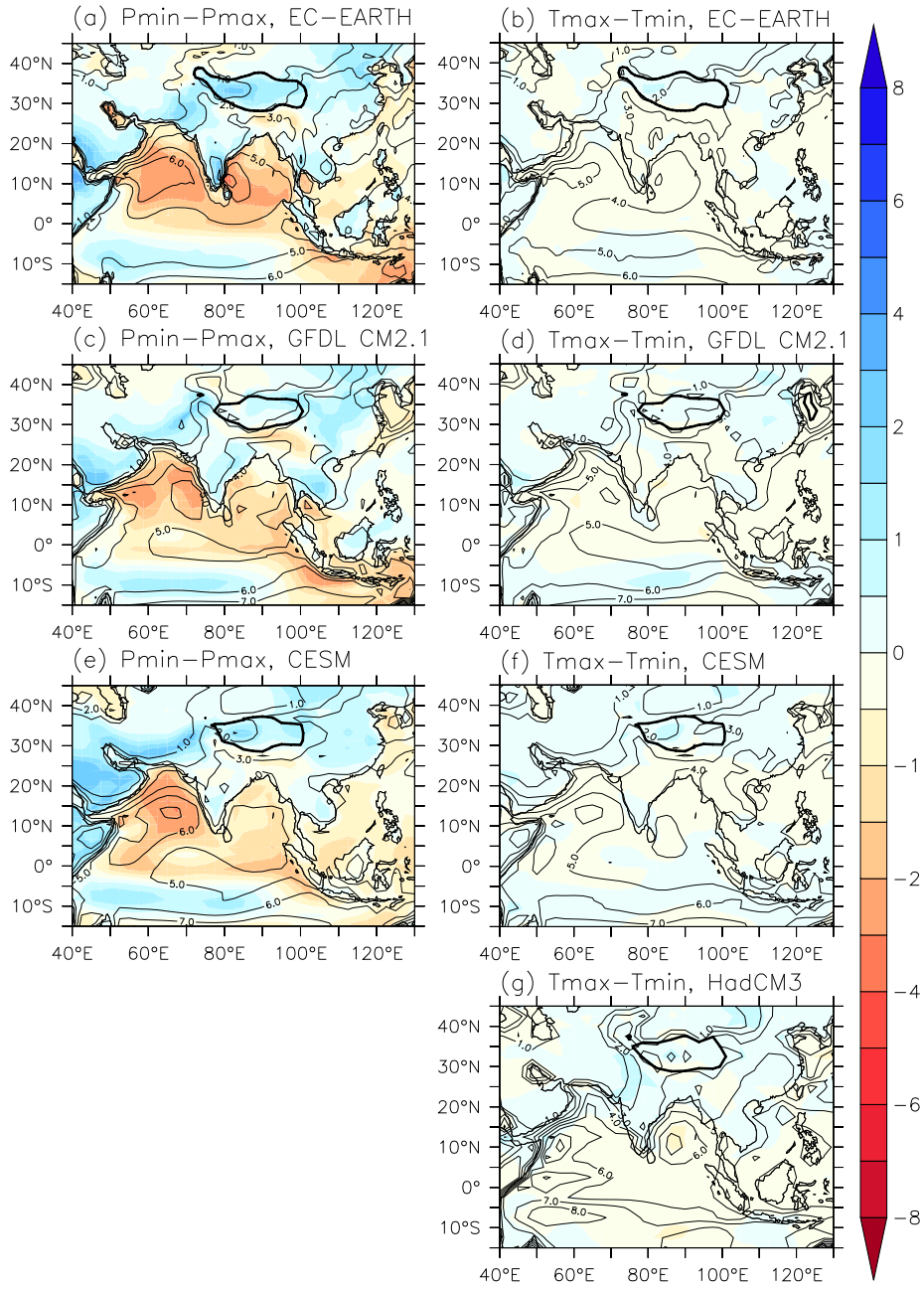


Figure C.8: June-July-August average evaporation difference in mm/day for $P_{min}-P_{max}$ (left) and $T_{max}-T_{min}$ (right) for all models. Contours indicate values for P_{max} (left) or T_{min} (right). The thick contour line is at 4km height, indicating the Tibetan Plateau. As in Figure 8 in the main text, but for all models.

GFDL CM2.1

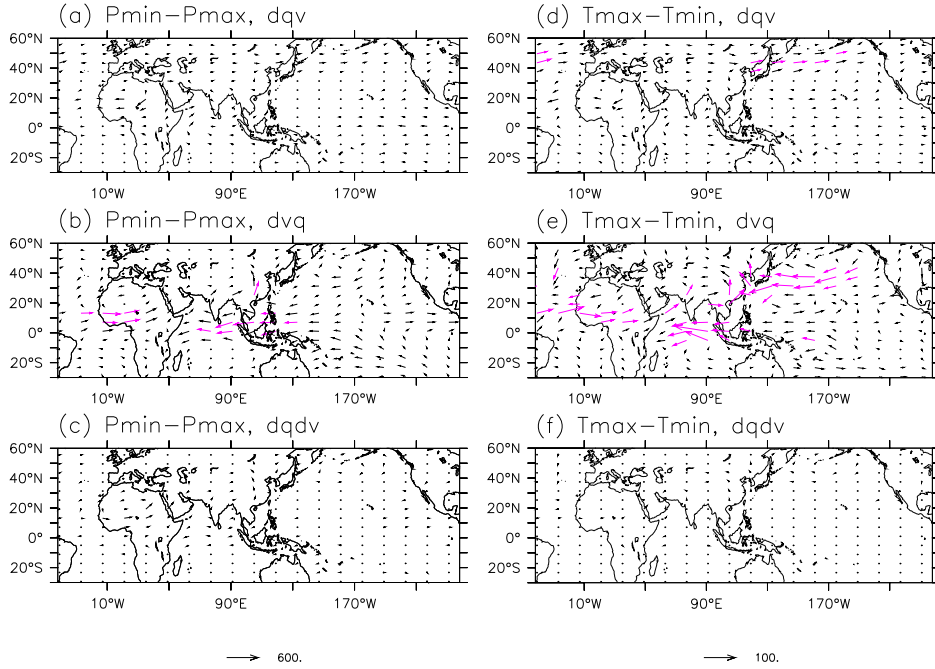


Figure C.10: Breakdown of June-July-August average moisture transport changes $d\mathbf{Q}$, the vertical integral of $d(q\mathbf{v})$ in $\text{kg}/(\text{ms})$, for precession ($P_{\min}-P_{\max}$, left) and obliquity ($T_{\max}-T_{\min}$, right) in GFDL. $d\mathbf{Q}$ is broken down following Equation 1 in (Bosmans et al., 2015a) into $dq\mathbf{v}$ (top, the thermodynamic part, related to changes in specific humidity, with $P_{\max} \mathbf{v}$), dvq (middle, the dynamic part, related to changes in wind, with $P_{\max} q$) and $dqdv$ (bottom, due to changes in both humidity and wind). Unit vector length for precession (left) is 600, with purple vectors indicating vectors larger than 200. For obliquity (right), vector length is 100, with purple vectors indicating changes in moisture transport larger than 30 $\text{kg}/(\text{ms})$. For these breakdown terms, monthly model output was used. The total moisture transport \mathbf{Q} for GFDL is given in Figures C.2, C.4.

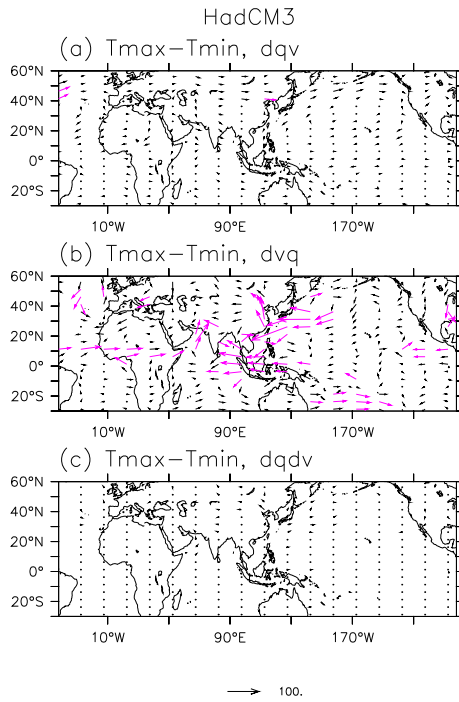


Figure C.12: Breakdown of June-July-August average moisture transport changes $d\mathbf{Q}$, the vertical integral of $d(\mathbf{q}\mathbf{v})$ in $\text{kg}/(\text{ms})$ for obliquity ($T_{\max} - T_{\min}$) in HadCM3. $d\mathbf{Q}$ is broken down following Equation 1 in (Bosmans et al., 2015a) into $d\mathbf{q}\mathbf{v}$ (top, the thermodynamic part, related to changes in specific humidity, with $P_{\max} \mathbf{v}$), $d\mathbf{v}\mathbf{q}$ (middle, the dynamic part, related to changes in wind, with $P_{\max} \mathbf{q}$) and $d\mathbf{q}\mathbf{d}\mathbf{v}$ (bottom, due to changes in both humidity and wind). Unit vector length for obliquity is 100, with purple vectors indicating changes in moisture transport larger than $30 \text{ kg}/(\text{ms})$. For these breakdown terms, monthly model output was used. The total moisture transport \mathbf{Q} for HadCM3 is given in Figure C.6.

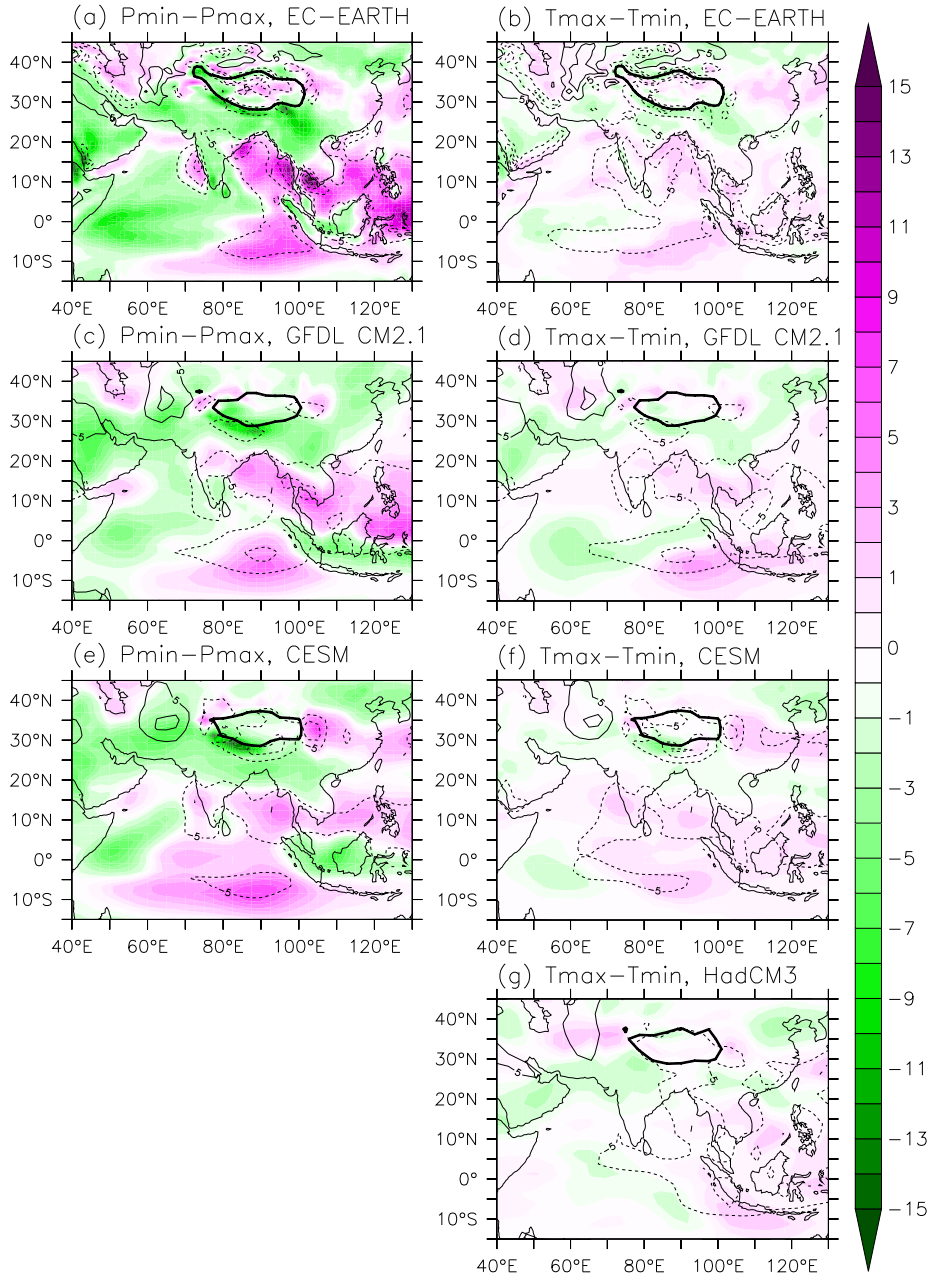


Figure C.13: June-July-August average vertical velocity at 500 hPa in 10^{-2}Pa/s for Pmin-Pmax (left) and Tmax-Tmin (right) for all models. Contours indicate values for Pmax (left) or Tmin (right) with negative values indicating upward motion. Green indicates more upward or less downward motion, purple indicates more downward or less upward motion. The thick contour line is at 4km height, indicating the Tibetan Plateau. As in Figure 9 in the main text, but for all models.

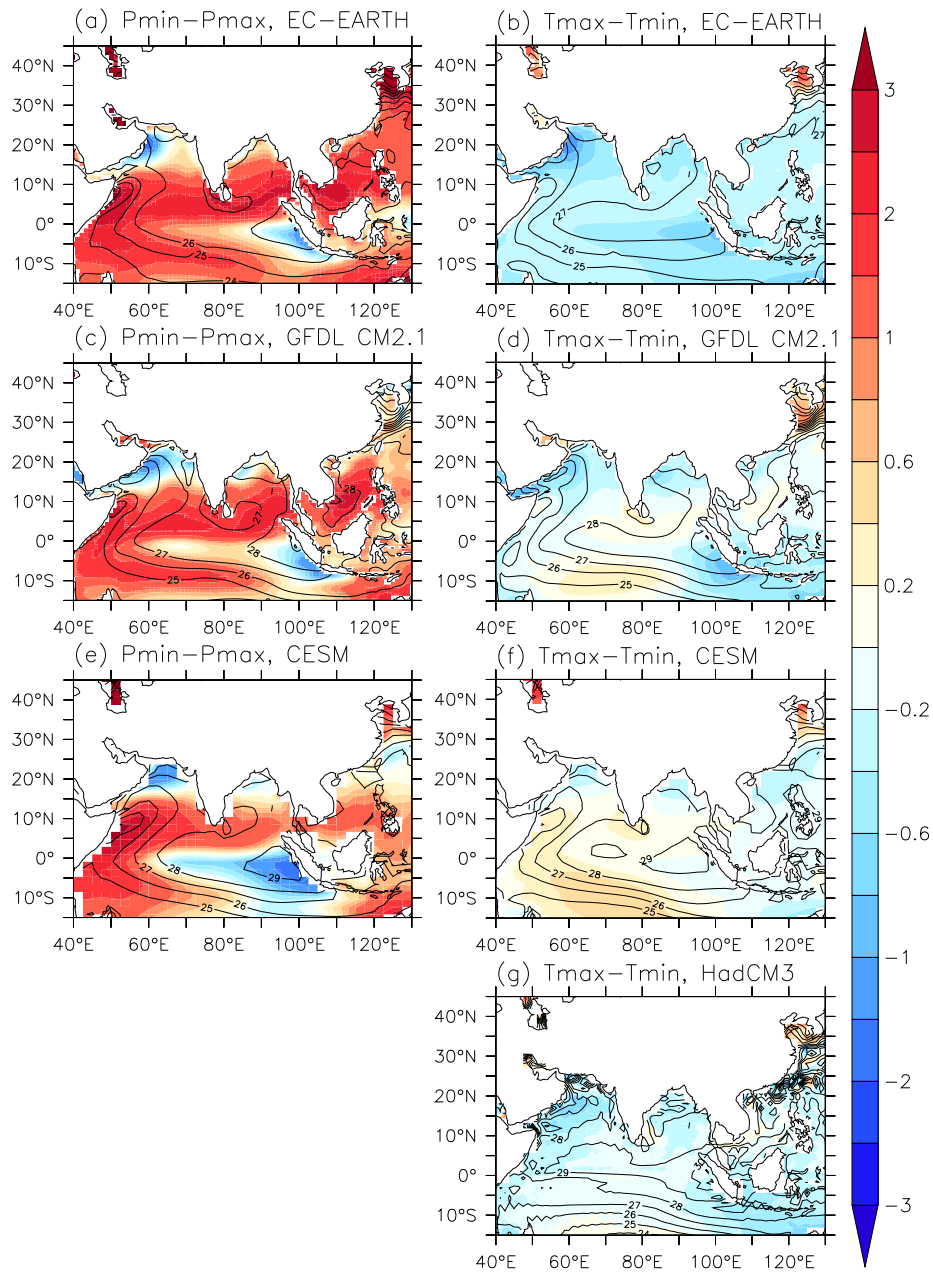


Figure C.14: June-July-August average sea surface temperature difference for $P_{min}-P_{max}$ (left) and $T_{max}-T_{min}$ (right) for all models. Contours indicate values for P_{max} (left) or T_{min} (right) in °C. As in Figure 10 in the main text, but for all models.