

*Northern Hemisphere monsoon response
to mid-Holocene orbital forcing and
greenhouse gas-induced global warming*

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1 **Northern Hemisphere monsoon response to**
2 **mid-Holocene orbital forcing and greenhouse**
3 **gas-induced global warming**

4 **Roberta D'Agostino¹**

5 **Jürgen Bader^{1,2}**

6 **Simona Bordoni³**

7 **David Ferreira⁴**

8 **Johann Jungclaus¹**

9 ¹Max Planck Institute for Meteorology, Hamburg, Germany.

10 ²Uni Climate, Uni Research and the Bjerknes Centre for Climate Research, Bergen, Norway.

11 ³California Institute of Technology, Pasadena, California.

12 ⁴Department of Meteorology, University of Reading, United Kingdom

13 **Key Points:**

- 14 • Different mechanisms mediate the response of Northern Hemisphere monsoons un-
15 der future global warming and mid-Holocene forcing.
- 16 • Northern Hemisphere monsoons intensify more strongly in mid-Holocene than in
17 future climate despite a larger warming in the latter.
- 18 • As an emergent constraint for future projections, tropical circulation weakening
19 limits monsoon rainfall increase with global warming.

Corresponding author: Roberta D'Agostino, Max Planck Institute for Meteorology, Bundesstr.
53, 20146, Hamburg, Germany, roberta.dagostino@mpimet.mpg.de

Abstract

Precipitation and circulation patterns of Northern Hemisphere monsoons are investigated in Coupled Model Intercomparison Project phase 5 simulations for mid-Holocene and future climate scenario rcp8.5. Although both climates exhibit Northern Hemisphere warming and enhanced inter-hemispheric thermal contrast in boreal summer, changes in the spatial extent and rainfall intensity in future climate are smaller than in mid-Holocene for all Northern Hemisphere monsoons except the Indian monsoon. A decomposition of the moisture budget in thermodynamic and dynamic contributions suggests that under future global warming the weaker response of the African, Indian and North American monsoons results from a compensation between both components. The dynamic component, primarily constrained by changes in net energy input over land, determines instead most of the mid-Holocene land monsoonal rainfall response.

1 Introduction

The mid-Holocene was a period around 6,000 years ago, when insolation changes driven by Earth's axis precession changes resulted in a general warming of the Northern Hemisphere (NH), an enhanced insolation seasonality and a stronger inter-hemispheric thermal contrast compared with present-day boreal summer (Zhao & Harrison, 2012). In agreement with expectations based on recent theories of monsoons (Schneider et al., 2014), these insolation-driven temperature changes resulted in a robust increase in monsoonal rainfall during the last interglacial and the mid-Holocene in North Africa (Weldeab et al., 2007; Tjallingii et al., 2008), India (Schulz et al., 1998; Fleitmann et al., 2003), East Asia (Liu & Ding, 1998; Yuan et al., 2004; Wang et al., 2008; Lézine et al., 2011; Hély & Lézine, 2014; Tierney & Pausata, 2017) and northernmost South America (Haug et al., 2001) as shown by proxy reconstructions. This wettening tendency is also observed in a number of climate simulations from the Paleoclimate Model Intercomparison Project (PMIP) (Zhao et al., 2005; Zhao & Harrison, 2012) under mid-Holocene forcing, despite difficulties in reproducing the magnitude and northward expansion of rainfall as suggested by proxy data, particularly over the Sahara (Braconnot et al., 1999; Liu et al., 2007; Braconnot et al., 2012; Harrison et al., 2015; Boos & Körtj, 2016). Some recent studies show better agreement with proxies on mid-Holocene precipitation in models that account for interactive vegetation or realistic vegetation cover over the Sahara (Vamborg et al., 2010; Swann et al., 2014; Pausata et al., 2016; Egerer et al., 2018; Lu et al., 2018), but sim-

52 ulations with precipitation and vegetation changes consistent with proxies have yet to
53 be achieved.

54 Similar to mid-Holocene, the Representative Concentration Pathway global warm-
55 ing scenario rcp8.5 projects a warming of the Northern relative to the Southern Hemi-
56 sphere and an enhanced inter-hemispheric thermal contrast resulting from stronger warm-
57 ing over land than over ocean (Sutton et al., 2007; Compo & Sardeshmukh, 2009; Jones
58 et al., 2013; Acosta Navarro et al., 2017). These elements all support a tendency toward
59 increased global monsoon rainfall strength and extent (Trenberth et al., 2000; Hsu et al.,
60 2012, 2013; Kitoh et al., 2013; Lee & Wang, 2014) associated with reinforced low-level
61 moisture convergence (Hsu et al., 2012; Kitoh et al., 2013; Lee & Wang, 2014). On a re-
62 gional scale, evaluation of Coupled Model Intercomparison Project phase 3 and 5 (CMIP3
63 and CMIP5) simulations has indicated a wettening of the Asian monsoon (Kitoh et al.,
64 2013; Endo & Kitoh, 2014) but has shown poor agreement in the African monsoon re-
65 gion because of competing effects of CO₂ increase and SST biases on the modelled West
66 African monsoon response (Biasutti, 2013; Gaetani et al., 2017). Projections of the North
67 American monsoon remain more inconclusive, with most models projecting a delay in
68 the monsoon season with no robust changes in its summer mean intensity (Cook & Sea-
69 ger, 2013; Seth et al., 2013, 2011). The extent to which this might be a result of exist-
70 ing biases in the simulations of the present-day monsoon climatology remains a topic of
71 debate (Pascale et al., 2017).

72 Despite a different global mean temperature response, the mean warming and the
73 enhanced inter-hemispheric temperature contrast would suggest a strengthening and widen-
74 ing of NH monsoons in both climates relative to pre-industrial conditions (Tab. S2). Nev-
75 ertheless, how similar the resulting regional monsoon responses are, remains unknown.

76 The energetic view of monsoons as moist energetically direct circulations tightly
77 connected to the global Hadley cell (Bordoni & Schneider, 2008; Schneider et al., 2014;
78 Biasutti et al., 2018) rather than as sea-breeze circulations driven by land-ocean tem-
79 perature contrast (Webster & Fasullo, 2002; Fasullo & Webster, 2003; Fasullo, 2012; Gadgil,
80 2018) might provide some insight into the differing response of NH monsoons to mid-
81 Holocene and rcp8.5 scenario. In this view, monsoons are fundamental components of
82 the tropical overturning circulation, and, like the global mean Hadley cell, they export
83 moist static energy (MSE) away from their ascending branches and precipitation max-

84 ima. If eddy energy fluxes are negligible, this implies that net energy input (NEI) into
85 the atmospheric column given by the difference between top-of-atmosphere radiative and
86 surface energy fluxes is primarily balanced by divergence of vertically integrated mean
87 MSE flux (Chou et al., 2001; Merlis et al., 2013; Boos & Korty, 2016, see Eq. (3) below).
88 Not surprisingly, the MSE budget has therefore provided the theoretical framework to
89 understand the response of monsoons to different surface heat capacity (i.e. ocean ver-
90 sus land) (Chou et al., 2001), changes in atmospheric dynamics (Tanaka et al., 2005; Vec-
91 chi & Soden, 2007), in tropical tropospheric stability (Neelin et al., 2003), and in veg-
92 etation (Kutzbach et al., 1996; Claussen & Gayler, 1997; Broström et al., 1998; Claussen
93 et al., 2013).

94 Changes in inter-hemispheric contrast in NEI, such as for instance those driven by
95 precession-induced insolation changes, require anomalous meridional energy transport
96 to restore energy balance. To the extent that during the summer most of this transport
97 is accomplished by monsoonal circulations (Heaviside & Czaja, 2013; Walker, 2017), this
98 would imply a shift of the monsoonal circulation ascending branches and precipitation
99 maxima into the hemisphere with increased NEI and, possibly, an associated circulation
100 strengthening (Schneider et al., 2014; Bischoff et al., 2017). It is important to note, how-
101 ever, that the MSE budget constrains the energy transport rather than the circulation
102 strength itself (Merlis et al., 2013; Hill et al., 2015). The degree to which changes in en-
103 ergy transport implied by a given radiative forcing are accomplished through just changes
104 in circulation strength or also changes in energy stratification (or gross moist stability,
105 Neelin and Held, 1987) is not fully understood.

106 Here, we investigate the NH monsoon response in CMIP5 simulations under rcp8.5
107 and mid-Holocene forcing factors. Given the stronger thermal contrast between hemi-
108 spheres and land versus ocean in rcp8.5 than in mid-Holocene one might expect that mon-
109 soon rainfall and extent would be greater in the former than in the latter. However, we
110 will show that the opposite is true. Mechanisms of this differing monsoon response are
111 investigated by decomposing the anomalous moisture budget in thermodynamic and dy-
112 namic components. The dynamic component is further related to NEI changes, to bet-
113 ter understand why monsoons respond differently to different climate forcings and to ex-
114 plore to what extent the mid-Holocene may be considered as an analogue of future green-
115 house gas-induced warming.

116 2 Data and Methods

117 We leverage mid-Holocene, piControl and rcp8.5 experiments that are available in
 118 CMIP5 archives. We use the first ensemble member (r1i1p1) of nine available models with
 119 all three experiments (i.e., bcc-csm-1-1, CCSM4, CNRM-CM5, CSIRO-Mk3-6-0, FGOALS-
 120 g2, HadGEM2-ES, IPSL-CM5A-LR, MIROC-ESM and MRI-CGCM3, see Table SI1).
 121 All datasets are interpolated to a common $1^\circ \times 1.25^\circ$ latitude/longitude grid and to 17
 122 pressure levels.

123 June to September (JJAS) climatologies are calculated for the last 30 years of rcp8.5,
 124 for the period 1850 - 2005 of piControl and for the last 100 years of mid-Holocene sim-
 125 ulations. September is also included in the summer season, to account for seasonality
 126 delays in the Hadley and monsoonal circulations in both mid-Holocene and rcp8.5 (Seth
 127 et al., 2010; Dwyer et al., 2012; Seth et al., 2013; D’Agostino et al., 2017).

128 Changes in monsoon extent and strength are assessed using the following metrics:
 129 the monsoon extent is the land-only area where annual precipitation range, defined as
 130 the difference between summer and winter rainfall, exceeds 2 mm/day for each monsoon
 131 domain. The selected threshold warrants a concentrated summer rainy season and dis-
 132 tinguishes monsoons from year-round rainy regimes (Zhou et al., 2008; Liu et al., 2009;
 133 Hsu et al., 2012). Choosing different definitions to calculate land-monsoon area (e.g. lo-
 134 cal summer precipitation exceeding 35%, 40%, 50% of the annual rainfall) does not sig-
 135 nificantly affect our results. The monsoon strength is the average summer rainfall cal-
 136 culated in each monsoon domain, specifically (see boxes in Fig. 1):

- 137 1. African monsoon (5° to 23.3° N, 20° W to 40° E).
- 138 2. Indian monsoon (5° to 23.3° N, 70° to 120° E).
- 139 3. North American monsoon (5° to 30° N, 120° W to 40° W).

140 We also consider the whole NH tropical land-monsoon area (NHM, 5° to 30° N,
 141 0 to 360° E). We exclude from our analyses the East Asian monsoon because its dynam-
 142 ics is related to shifts of the Pacific Subtropical High and interactions between the jet-
 143 stream and the Asian topography rather than to ITCZ seasonal migration and regional
 144 Hadley cell dynamics (Chen & Bordoni, 2014; Zhisheng et al., 2015). Following Trenberth
 145 and Guillemot (1995), the linearized anomalous moisture budget is decomposed into ther-
 146 modynamic, dynamic components and a residual (*Res*) as:

$$\rho_w g \delta(P - E) = - \int_0^{P_s} \nabla \cdot (\delta \bar{q} \bar{\mathbf{u}}_{\text{piControl}}) dp - \int_0^{P_s} \nabla \cdot (\bar{q}_{\text{piControl}} \delta \bar{\mathbf{u}}) dp - Res, \quad (1)$$

147 where overbars indicate monthly means, $(P-E)$ is precipitation minus evapora-
 148 tion, p is pressure, q is specific humidity, $\bar{\mathbf{u}}$ is the horizontal vector wind, and ρ_w is the
 149 water density. δ indicates the difference between each experiment (mid-Holocene or rcp8.5)
 150 and the reference climate (piControl) as:

$$\delta(\cdot) = (\cdot)_{\text{mid-Holocene or rcp8.5}} - (\cdot)_{\text{piControl}}. \quad (2)$$

151 In Eq. (1), the first term on the right-hand side is the thermodynamic contribu-
 152 tion (TH): it represents changes in moisture flux convergence arising from changes in mois-
 153 ture, which generally follow the Clausius-Clapeyron relation for negligible relative hu-
 154 midity changes (e.g. Held and Soden, 2006). The second term in Eq. (1), the dynamic
 155 contribution (DY), involves changes in winds with unchanged moisture, and is mostly
 156 related to changes in the mean atmospheric flow. The third term describes the residual
 157 (Res) which accounts for transient eddy contribution and surface quantities as described
 158 in the Supplementary Information.

159 Changes in the DY contribution to monsoonal precipitation changes are related to
 160 patterns of anomalous NEI, as any anomalous NEI in monsoonal regions will require changes
 161 in MSE export by the mean circulation in steady state:

$$\nabla \cdot \{\overline{\mathbf{u}h}\} = NEI = R_{TOA} - F_{sfc}, \quad (3)$$

162 where $\{\overline{\mathbf{u}h}\}$ is the vertically integrated MSE flux, R_{TOA} the net top-of-atmosphere ra-
 163 diative fluxes and F_{sfc} the sum of the surface radiative and turbulent enthalpy fluxes.

164 **3 Results**

165 The future rcp8.5 and the past mid-Holocene climates are associated, respectively,
 166 with a strong (+4.2 K) and a weak (+0.3 K) global warming signal relative to piCon-
 167 trol (Fig. 1, upper panels; Table S2). They also exhibit higher inter-hemispheric ther-
 168 mal contrasts (+10.0 K and +9.7 K compared to +9.2 K for piControl, see Table S2).
 169 However, the precipitation difference between rcp8.5 and mid-Holocene (Fig. 1, lower panel)

170 reveals a complex pattern of relative drying and wettening, reflective of a general ten-
171 dency towards land drying and ocean wettening in rcp8.5, and land wettening and ocean
172 drying in mid-Holocene.

173 To explain these differences in the precipitation response, we analyze the anoma-
174 lous moisture budget of the two climates relative to piControl. This analysis shows how
175 changes in net precipitation $\delta(P-E)$ (see Eq. (1)) are primarily due to changes in pre-
176 cipitation alone, with changes in evaporation being negligible both in the multi-model
177 mean (Figure S1 and S2) and in each individual model (not shown). Relative to piCon-
178 trol, precipitation in the African and Indian monsoons generally increases in mid-Holocene,
179 while it decreases in the North American monsoon and increases in the Indian monsoon
180 in rcp8.5. Figure 2 shows a general wettening of African and Indian monsoons in mid-
181 Holocene relative to piControl, while in rcp8.5 the North American monsoon dries and
182 the Indian monsoon wettens. The drying in the North American monsoon seen under
183 rcp8.5 in the models considered in this study is at odds with previously published stud-
184 ies, which suggest no robust changes in the mean monsoon precipitation, but is in agree-
185 ment with simulations in which SST biases in the North Atlantic are corrected with flux
186 adjustment (Pascale et al., 2017). These ensemble mean ($P-E$) changes are robust as
187 they occur in at least 8 out of 9 models considered here (stippled areas in Fig. 2), but mod-
188 els disagree on the magnitude of these changes. However, while in mid-Holocene mod-
189 els robustly produce wettening in the African equatorial rain belt and the sub-Saharan
190 region, particularly in those models with active land module (i.e. bcc-csm1-1, CCSM4,
191 CNRM-CM5, IPSL-CM5A-LR, FGOALS-g2, Had-GEM-ES, MIROC-ESM), there is less
192 consensus on net precipitation changes in rcp8.5. Only CCSM4 shows a wettening of equa-
193 torial Africa; other models show decreased or no change in monsoonal precipitation (not
194 shown).

195 It is noteworthy that, on a global scale (including changes over land as well as over
196 oceans), rcp8.5 exhibits a robust shift of tropical precipitation towards the near-equatorial
197 ocean relative to piControl (Fig. 2c). This tendency is also consistent with the projected
198 squeezing of rain belts around the equator and the narrowing of the ITCZ in rcp8.5 (Byrne
199 & Schneider, 2016). These findings however highlight that global ITCZ changes are not
200 a good indicator of the land monsoon changes.

201 It is readily apparent from Figs. 1 and 2 that the mid-Holocene monsoon response
 202 is not a weaker version of the rcp8.5 response. Even more surprisingly, the simulated land
 203 monsoon changes are almost systematically smaller in rcp8.5 than in mid-Holocene, de-
 204 spite stronger global mean temperature increase and a slightly larger inter-hemispheric
 205 thermal contrast in the former than in the latter. In fact, both extent and strength of
 206 individual monsoons and the global NH land monsoon are projected to increase more
 207 in mid-Holocene than in rcp8.5. The notable exception to this general pattern is the In-
 208 dian monsoon, whose strength increases more in rcp8.5 (Tab. 1).

209 To explain why the monsoon response is weaker under future global warming rel-
 210 ative to the mid-Holocene, we decompose $\delta(P-E)$ in TH and DY contributions as de-
 211 scribed in Section 2. Each of these components is shown in Figure S3 and S4; Results
 212 are summarized in Fig. 3 by averaging these components in each monsoon domain, where
 213 annual-range precipitation exceed 2 mm/day. The magnitude of the residual relative to
 214 the other components is also shown.

215 Fig. 3 reveals a striking contrast in the response in the two climates: in mid-Holocene,
 216 the DY term dominates the anomalous moisture budget in the African and Indian mon-
 217 soon regions and in the overall NH monsoon domain. Only in the North American mon-
 218 soon region does this term contribute marginally to the anomalous moisture budget (Fig.
 219 3, and Fig. S3b). The DY component increases NH land precipitation through increased
 220 moisture convergence there (Fig. S3; see methods in Supplementary Information). Like-
 221 wise, drying over near-equatorial oceans is associated with weaker wind convergence, es-
 222 pecially in the Atlantic sector. Therefore, the enhanced African and Indian monsoonal
 223 rainfall in mid-Holocene is due to a strengthening of the mean flow. On the other hand,
 224 the TH component plays a secondary role in the mid-Holocene net precipitation increase
 225 in all monsoon domains, except in the North American monsoon (Fig. 3a and Fig. S3
 226 a and c). On average, the TH and DY terms tend to reinforce each other, both contribut-
 227 ing to a wettening tendency.

228 In contrast, the overall weaker wettening in future rcp8.5 projections results from
 229 a compensation between the DY term and the TH term (where the latter moistens mon-
 230 soons as the climate warms) (Fig. 3b and Fig. S4). The substantial drying of the North
 231 American monsoon arises mainly from a strong weakening of the mean circulation (DY
 232 term, Table 1). On the other hand, the TH and the DY components feature strong spa-

233 tial variations in the Indian monsoon region: the TH plays a major role in the wettening
234 tendency over the eastern Indian peninsula, and is responsible for the strong drying
235 on its western part (Fig. S4). However, averaging over the entire domain, the TH
236 term dominates over the DY term, and drives an overall wettening.

237 These analyses suggest therefore that the wettening and northward shift of NH mon-
238 soons in mid-Holocene arises mainly from the strengthening of the mean circulation. On
239 the other hand, the weak monsoon response to anthropogenic forcing in rcp8.5 relative
240 to mid-Holocene is mainly due to a compensation between the thermodynamically driven
241 wettening and a dynamically driven drying, as already pointed out by some previous stud-
242 ies (Seager et al., 2010, 2014; Endo & Kitoh, 2014).

243 Tropical circulation weakening with warming (i.e. weakening of the DY component
244 in all considered monsoons) is a consequence of increased stability in the tropics where
245 temperature lapse rates follow moist adiabats (Held & Soden, 2006). Over tropical and
246 subtropical continents, the stability increase is not compensated by increases in low-level
247 MSE which reduces convection and moisture convergence from oceans, with an associ-
248 ated reduction in land monsoonal rainfall (Fasullo, 2012). The projected monsoonal cir-
249 culation weakening relative to mid-Holocene hence represents a constraint for monsoonal
250 rainfall: precipitation squeezes around the tropical ocean in rcp8.5 as the static stabil-
251 ity increases, the circulation weakens and continental moisture convergence decreases.
252 Unlike what is seen in rcp8.5, the strengthening of the circulation in mid-Holocene al-
253 lows for increased moisture convergence over land monsoon regions, with a shift of the
254 tropical precipitation from ocean to land and stronger monsoonal rainfall than projected
255 in rcp8.5.

256 To further understand, at least qualitatively, the different response of land-ocean
257 monsoonal rainfall in the two climates, we analyze changes in NEI in mid-Holocene and
258 rcp8.5 relative to piControl (Fig. 3 and 4). In mid-Holocene, the NEI response is mainly
259 positive over NH continents relative to piControl primarily because of precession-induced
260 insolation changes (Fig. 4a). On the other hand, patterns of anomalous NEI are of op-
261 posite sign in rcp8.5, with positive values over the tropical ocean. Hence, to compensate
262 for these NEI changes, the mid-Holocene atmospheric circulation needs to export more
263 energy away from land regions, through a strengthening of the associated DY term (Fig. 3).
264 In rcp8.5, increased stability and the absence of such energetic forcing over NH lands,

265 where the energy budget is controlled by the top of the atmosphere radiation due to the
266 small thermal inertia of land (Neelin & Held, 1987), cause a weakening of the monsoonal
267 circulation an overall decrease of tropical land rainfall relative to mid-Holocene. Fig. 3
268 shows in fact a systematic NEI increase of $\sim 8 \text{ W/m}^2$ in mid-Holocene, compared to a
269 weak change ($< 1 \text{ W/m}^2$) in rcp8.5.

270 4 Discussion and Conclusions

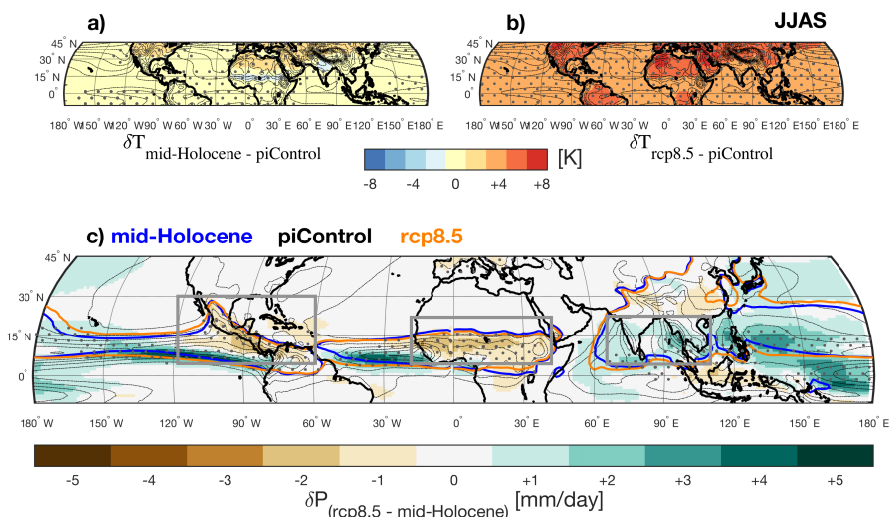
271 Here, we have investigated mechanisms of monsoon moistening and expansion in
272 two climates, mid-Holocene and future climate scenario rcp8.5. In both climates, the sim-
273 ulated NH summer monsoon rainfall is stronger and monsoon area wider than in the pre-
274 industrial era. However, the projected monsoon response to global warming is weaker
275 than in the simulated past, despite a much larger global warming in the former than in
276 the latter.

277 In rcp8.5, the NH land monsoon is expected to become wetter relative to pre-industrial
278 conditions because the atmospheric specific humidity increase leads to enhanced precip-
279 itation (thermodynamic effect). Additionally, the Hadley circulation is projected to ex-
280 pand and weaken in the future (Frierson et al., 2007; Lu et al., 2007; Seidel et al., 2008;
281 D’Agostino et al., 2017) following the widening and the slowdown already observed in
282 recent decades (Hu & Fu, 2007; Birner, 2010; Davis & Rosenlof, 2012; Nguyen et al., 2013;
283 D’Agostino & Lionello, 2017). This weakens the dynamic term of the moisture budget.
284 Therefore, the weak monsoonal rainfall response with global warming generally results
285 from a compensation between the thermodynamic and dynamic terms. The degree of
286 compensation differs strongly among monsoon regions. For instance, in the Indian mon-
287 soon the TH component overwhelms the DY component, giving rise to an overall wet-
288 tening; in the North American monsoon, the DY component is dominant and respon-
289 sible for a significant drying.

290 Unlike what happens under greenhouse gas-induced warming, the strengthening
291 of the mean atmospheric flow is the dominant mechanism behind the wettening and widen-
292 ing of NH monsoons in mid-Holocene. The circulation brings more rainfall over land than
293 over ocean, expanding the total NH land-monsoon area further northward than in rcp8.5.
294 In fact, the dynamic response reinforces the thermodynamically driven wettening in mid-
295 Holocene; in contrast the two components partially cancel each other in rcp8.5.

296 Advances in our theoretical understanding of monsoons allows us to link dynamically-
297 induced precipitation changes to changes in NEI (Chou et al., 2001; Neelin et al., 2003;
298 Byrne & Schneider, 2016). In this framework, monsoonal circulations, as part of the global
299 tropical overturning, export MSE away from their ascending branches. In steady state,
300 the net MSE flux divergence balances the NEI. Therefore to the extent that energy strat-
301 ification does not change significantly, changes in NEI need to be compensated for by
302 changes in circulation strength. Hence, the different monsoon responses in the two cli-
303 mates can ultimately be related to changes in the forcing itself, which influences differ-
304 ently the NEI over land and over ocean. In fact, the shortwave forcing, which dominates
305 the mid-Holocene, exhibits a stronger land-ocean contrast than the longwave perturba-
306 tion associated with greenhouse gas increases in rcp8.5 (Fig. S5). In mid-Holocene, the
307 stronger cross-equatorial atmospheric circulation and the enhanced dynamic term are
308 a result of increased energetic input over the continents: the atmospheric circulation must
309 be stronger in order to export energy away from these regions in the past climate. The
310 absence of such energetic forcing over NH lands in rcp8.5 relative to mid-Holocene re-
311 sults in a relative weakening of mean circulation and hence of the associated precipita-
312 tion. The strengthening of the dynamic component, therefore, represents a key ingre-
313 dient for monsoon widening and wetting in mid-Holocene. The weakening of the trop-
314 ical circulation with global warming limits the projected expansion and intensification
315 of the monsoon systems. The degree of compensation between the thermodynamic and
316 dynamic responses with warming remains highly uncertain and might contribute signif-
317 icantly to the inter-model spread in CMIP5 simulations (Stocker et al., 2014).

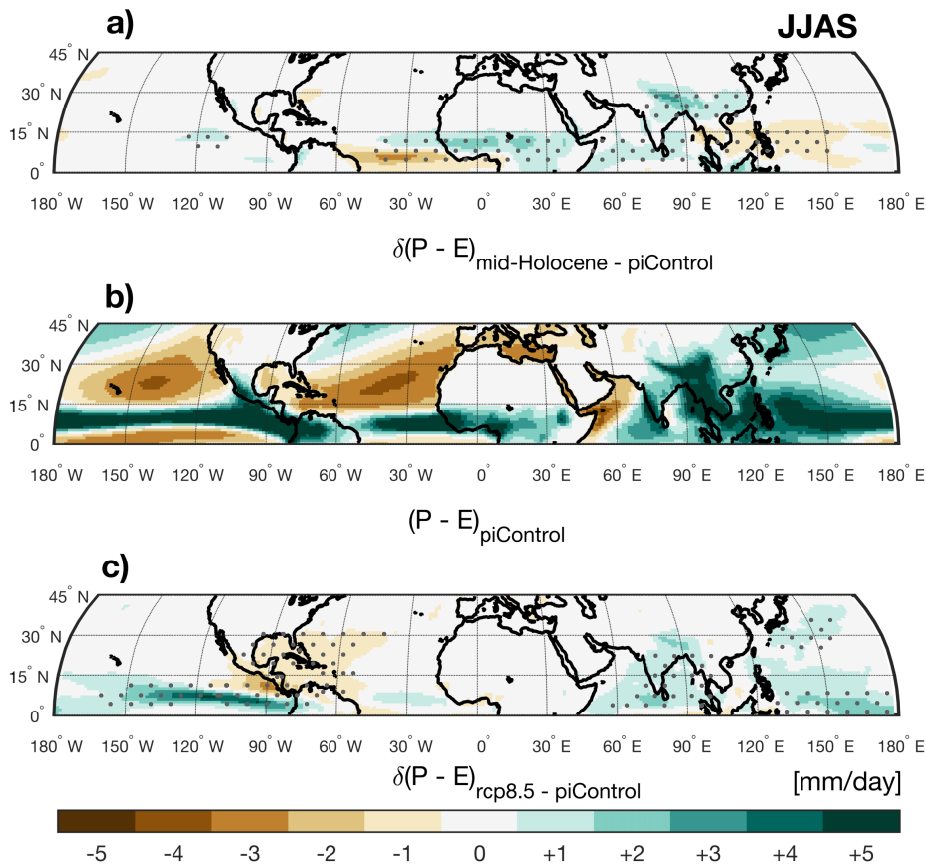
318 This process-oriented study takes an important step towards improving our under-
319 standing of monsoon dynamics, quantifying the important role of atmospheric circula-
320 tion changes in monsoonal precipitation changes by comparing and contrasting past and
321 future climates. Our results highlight that mean surface warming and inter-hemispheric
322 contrast in surface warming are poor indicators of the monsoonal precipitation response.
323 Rather, the monsoon response is constrained by the integrated energy balance, which
324 accounts for changes at the surface as well as at the top of the atmosphere. This explains
325 why the mid-Holocene does not represent an analogue for future warming.



326 **Figure 1.** Surface temperature difference between mid-Holocene (a) and rcp8.5 (b) and
 327 piControl in June-to-September (JJAS) ensemble means (shading). Precipitation difference be-
 328 tween rcp8.5 and mid-Holocene JJAS ensemble means (c, shading). Black dashed lines in every
 329 panel show the piControl as reference (contour interval 2 K for temperature and 2 mm/day for
 330 precipitation). Orange and blue bold lines in c) show areas within which the annual precipitation
 331 range (JJAS minus DJFM) exceeds 2 mm/day for rcp8.5 and mid-Holocene, respectively. Grey
 332 boxes indicate the North American, African and Indian monsoon domains. Stippling indicates
 333 areas where at least 8 out of 9 models agree on the sign of the change.

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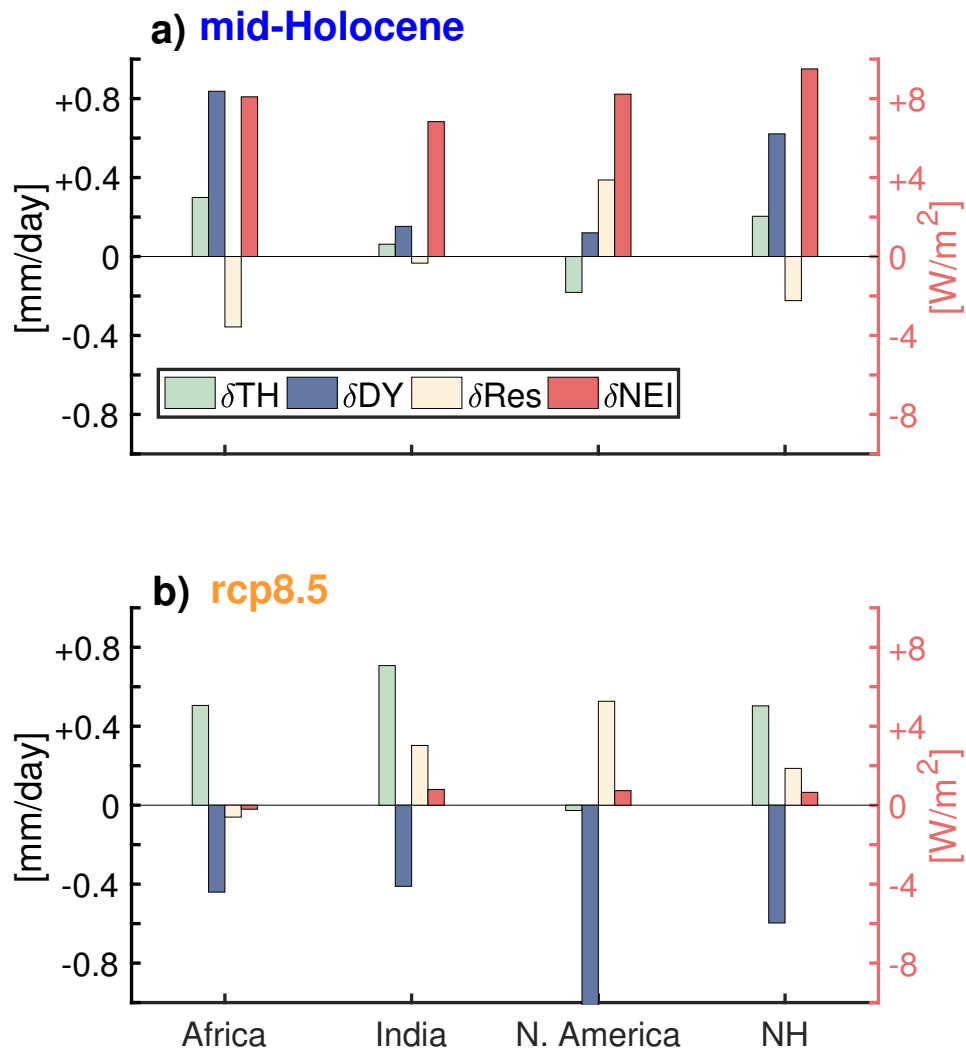
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 359 [data.dkrz.de/search/cmip5-dkrz/](https://esgf-data.dkrz.de/search/cmip5-dkrz/). Scripts used in the analysis and other supporting in-
 360 formation useful to reproduce the author's work are archived by the Max Planck Insti-
 361 tute for Meteorology and can be obtained contacting: publications@mpimet.mpg.de.



334 **Figure 2.** Net precipitation difference between the mid-Holocene (a) and the rcp8.5 (c) rel-
 335 ative to *piControl* in June-to-September (JJAS) ensemble means (shading). PiControl is also
 336 shown as reference (b). Black dashed lines in each panel show the piControl as reference (contour
 337 interval 20 W/m²). Stippling indicates areas where at least where 8 out of 9 models agree on the
 338 sign of the change.

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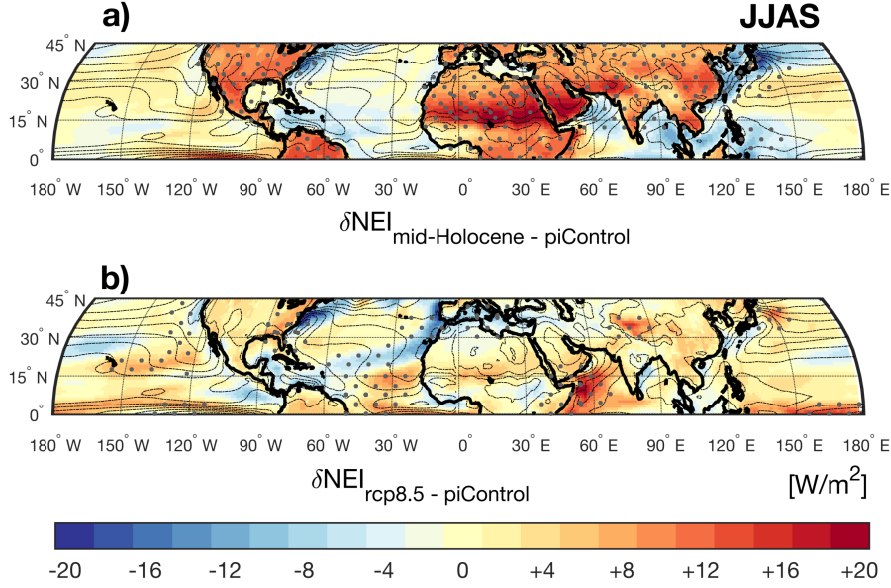


339 **Figure 3.** Regionally averaged Net Energy Input (NEI - red axis) changes and changes in
 340 thermodynamic (δTH) and dynamic (δDY) components of the moisture budget, as well as its
 341 residual (δRes) (see Eq.1) for mid-Holocene (a) and rcp8.5 (b) (black axis). Note that 8 out of 9
 342 models agree on the sign of the change.

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343 **Figure 4.** Net energy input (NEI) difference between mid-Holocene (a) and rcp8.5 (b) relative
 344 to piControl in June-to-September (JJAS) ensemble means (shading). Stippling indicates areas
 345 where at least 8 out of 9 models agree on the sign of the change.

346 **Table 1.** Changes in mid-Holocene and rcp8.5 land monsoon extent and strength relative to
 347 piControl. Standard errors for piControl models are reported in brackets. The monsoon extent is
 348 calculated inside each monsoon domain where the difference between JJAS and DJFM precipita-
 349 tion exceeds 2 mm/day, as shown in solid lines in Fig. 1.

Monsoons	Extent (10° Km)			Strength (mm/day)			ITCZ (lat. degs)			$\phi_{Pr>2mm/day}$		
	piControl	mid-Holocene	rcp8.5	piControl	mid-Holocene	rcp8.5	piControl	mid-Holocene	rcp8.5	piControl	mid-Holocene	rcp8.5
African	5.2 (± 0.7)	+15.4%	+4.4%	5.3 (± 11.0)	+20.3%	+1.2%	7.5	8.4	7.5	14.2	15.4	14.2
Indian	3.1 (± 0.4)	+9.2%	+7.4%	8.5 (± 1.3)	+1.6%	+4.8%	11.5	11.6	11.4	21.8	22.9	22.6
North American	2.8 (± 0.5)	+3.7%	-4.3%	5.8 (± 1.3)	+7.8%	-5.8%	8.1	8.3	7.7	20.1	20.0	21.6
NH	9.3 (± 1.0)	+15.1%	+4.8%	7.0 (± 0.5)	+1.1%	-1.8%	7.9	7.9	7.2	19.2	19.6	18.2

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