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1 **Understanding the mechanisms behind the northward extension of the West African**
2 **Monsoon during the Mid-Holocene**

3

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23 **Abstract**

24

25 Understanding the West African monsoon (WAM) dynamics in the mid-Holocene (MH) is a
26 crucial issue in climate modelling, because numerical models typically fail to reproduce the
27 extensive precipitation suggested by proxy evidence. This discrepancy may be largely due to
28 the assumption of both unrealistic land surface cover and atmospheric aerosol concentration.
29 In this study, the MH environment is simulated in numerical experiments by imposing
30 extensive vegetation over the Sahara and the consequent reduction in airborne dust
31 concentration. A dramatic increase in precipitation is simulated across the whole of West
32 Africa, up to the Mediterranean coast. This precipitation response is in better agreement with
33 proxy data, in comparison with the case in which only changes in orbital forcing are
34 considered. Results show a substantial modification of the monsoonal circulation,
35 characterized by an intensification of large-scale deep convection through the entire Sahara,
36 and a weakening and northward shift ($\sim 6.5^\circ$) of the African easterly jet. The greening of the
37 Sahara also leads to a substantial reduction in African easterly wave activity and the
38 associated precipitation. The reorganization of the regional atmospheric circulation is driven
39 by the vegetation effect on radiative forcing and associated heat fluxes, with the reduction in
40 dust concentration to enhance this response. The results for the WAM in the MH present
41 important implications for understanding future climate scenarios in the region and in
42 teleconnected areas, in the context of projected wetter conditions in West Africa.

43

44 **Key words:** African humid period, African easterly waves, dust, land cover, Green Sahara,
45 EC-Earth

46 **1 Introduction**

47

48 Understanding climate variability in West Africa is one of the most challenging issues in
49 current climate science, from both the academic and socio-economic perspectives. Indeed,
50 food security in West African countries is largely based on rain fed agriculture [IFAD, 2013]
51 and is potentially vulnerable to climate variability [Sultan and Gaetani, 2016]. Precipitation in
52 West Africa is concentrated in the boreal summer and is primarily associated with the West
53 African Monsoon (WAM) [Nicholson, 2013]. The WAM underwent large variability in the
54 20th century, showing a succession of long lasting wet and dry periods following
55 multidecadal trends [Rodríguez-Fonseca et al., 2015]. In this context, the need for reliable
56 predictions of rainfall in West Africa on decadal and longer time scales is crucial. However,
57 state-of-the-art climate models are still deficient in correctly reproducing the historical
58 variability and agreeing on future projections [Biasutti, 2013]. The skill in identifying and
59 disentangling the numerous drivers of the WAM variability that operate at different spatial
60 and temporal scales is modest at best [Biasutti et al., 2009; Rowell, 2013].

61

62 Furthermore, climate change in West Africa has been much larger during the Holocene
63 interglacial period than in the 20th century. Extreme droughts, developing in a matter of few
64 years or decades, have been recorded during late Holocene (5-0 kyr BP), with dramatic lake
65 low stands near the Equator [Shanahan et al., 2006] and the complete drying of fresh water
66 bodies in the Sahel [Gasse, 2000]. The opposite scenario occurred during early Holocene,
67 between 15 and 5 kyr ago, when increased summer precipitation led to an expansion of the
68 North African lakes and wetlands and an extension of grassland and shrubland into areas that
69 are now desert, giving origin to the so-called “Green Sahara”, or African Humid Period
70 (AHP) [Holmes, 2008]. Model simulations for the mid-Holocene (MH, 6 kyr BP) show

71 limitations in reproducing the intensification and geographical expansion of the WAM
72 precipitation [Harrison et al., 2014], and these shortcomings may be similar to those affecting
73 historical simulations and future projections. In general, numerical simulations of the WAM
74 dynamics are particularly sensitive to the shortcomings commonly affecting climate model
75 physics, such as the inaccurate representation of clouds, energy fluxes, coupling at the
76 surface, and convection at the sub-grid scale. Moreover, the coarse resolution of global
77 climate models is insufficient to properly simulate the mesoscale convection characterizing
78 the WAM precipitation. In historical simulations, fully coupled models exhibit biases in
79 reproducing the tropical Atlantic dynamics, resulting in warmer SST and weaker monsoonal
80 circulation [Roehrig et al., 2013]. In the description of future climate scenarios, the model
81 sensitivity to the global surface warming, and to the direct CO₂ radiative forcing, appears
82 crucial for the sign of the projected precipitation trends [Giannini, 2010; Gaetani et al.,
83 2016].

84

85 A number of studies have addressed the limitations in modelling the West African climate in
86 the MH, pointing to Saharan land cover as the key ingredient to simulate precipitation
87 patterns fitting paleoclimate data. By using a coupled atmosphere-vegetation model, Claussen
88 and Gayler [1997] found an amplification of the response of the atmospheric circulation to
89 changes in the Earth's orbit produced by a positive precipitation-vegetation feedback. By
90 prescribing a vegetated Sahara, several authors found a good agreement between simulated
91 precipitation and paleoclimate data, and identified the increased low-level moisture, which
92 destabilizes the vertical column and enhances convection, as a crucial factor to maintain the
93 intensification of the monsoon system due to the orbital forcing [Su and Neelin, 2005;
94 Patricola and Cook, 2007; Swann et al., 2014]. The use of interactive vegetation climate
95 models confirms the positive precipitation-vegetation feedback and the key role of the latent

96 heat increase in favoring the convection [Patricola and Cook, 2008; Rachmayani et al., 2015],
97 though the importance of the decrease in surface albedo is also highlighted [Levis et al.,
98 2004]. A recent study has shown that accounting for the large reduction in airborne dust
99 emission is also essential for an accurate simulation of the WAM strength during the MH
100 [Pausata et al., 2016]. The authors pointed out that the Paleoclimate Modelling
101 Intercomparison Project (PMIP) [Braconnot et al., 2011] and the Coupled Model
102 Intercomparison Project (CMIP) [Taylor et al., 2012] protocols specify the MH dust
103 concentration and land cover to be equal to the pre-industrial (PI) period, which may be the
104 main cause of the models' dry bias in simulating the MH WAM. In recent numerical
105 experiments, the inclusion of both dynamic vegetation and dust emissions is not sufficient to
106 reproduce an intensification of the WAM compatible with paleoclimate data, if the initial
107 configuration for the vegetation is set to PI levels. Indeed, the simulated precipitation changes
108 do not allow vegetation to grow north of 15° N, and the associated variations in dust load
109 compared to PI climate are limited, and so are the feedbacks [Harrison et al., 2015].

110

111 Building upon the study of Pausata et al. [2016], this work aims to describe in detail the
112 dynamics of the WAM during the MH, elucidating the mechanisms whereby changes in land
113 cover and associated mineral dust emission alter atmospheric circulation and precipitation
114 patterns. For this purpose, fully coupled climate model experiments, in which Saharan
115 vegetation and dust concentrations are changed concomitantly and in turn, are analyzed. In
116 the context of the existing literature on the topic, the analysis of the mineral dust's role in
117 modifying the atmospheric dynamics under vegetated Sahara conditions represents a novel
118 and original contribution to the understanding of the WAM system in the MH. The results of
119 this study are particularly relevant not only from a paleoclimate perspective but also with
120 respect to the assessment of future scenarios for West Africa [Schmidt et al., 2014]. Indeed, a

121 number of projections simulate a positive precipitation trend, and a subsequent land greening
122 and dust reduction [Evan et al., 2016]. The paper is organized as follows: the climate model
123 and the experimental design are described in section 2, the results are presented and discussed
124 in section 3, and the main findings are summarized in section 4.

125

126 **2 Model and experimental setup**

127

128 In the present study, the climate simulations performed by Pausata et al. [2016], using version
129 3.1 of the atmosphere-ocean fully coupled climate model EC-Earth [Hazeleger et al., 2010],
130 are analyzed. The atmospheric model is based on the Integrated Forecast System (IFS cycle
131 36r4) [ECMWF, 2010], including the H-TESEL land model [van den Hurk et al., 2000]. In
132 the land model, each grid-box is divided into fractions (bare ground, low and high vegetation,
133 intercepted water, shaded and exposed snow are represented over land), with properties
134 defining separate heat and water fluxes used in an energy balance equation solved for the skin
135 temperature. A fraction of the rainfall is collected by an interception layer, where the
136 remaining fraction is partitioned between surface runoff and infiltration. Infiltration and
137 runoff schemes depend on the soil texture and standard deviation of orography. The
138 simulation is run at T159 horizontal spectral resolution ($\sim 1.125^\circ$, approximately 125 km)
139 with 62 vertical levels. The atmospheric component is coupled by the OASIS 3 coupler
140 [Valcke, 2006] to the Nucleus for European Modelling of the Ocean version 2 (NEMO)
141 [Madec, 2008], and the Louvain-la-Neuve Sea Ice Model version 3 (LIM3) [Vancoppenolle
142 et al., 2008]. The ocean component NEMO has a nominal horizontal resolution of 1° and 46
143 vertical levels.

144

145 A control simulation (CTL) is performed following the PMIP3 protocol [Braconnot et al.,
146 2011]. The orbital forcing is set to MH values (6 kyr BP). For the greenhouse gases, the
147 methane is set at 650 ppb, instead of the 760 ppb PI value, and CO₂ and other greenhouse
148 gases are set to PI levels. Aerosol particle emissions, the solar constant, ice sheets,
149 topography and coastlines are set to PI conditions. The CTL run is compared to three
150 sensitivity experiments, which simulate the MH climate in the presence of a Green Sahara
151 with reduced atmospheric dust concentrations (GSRD), a Green Sahara with PI dust (GSPD),
152 and a present-day Sahara with reduced dust (PSRD).

153

154 The changes in vegetation in the GSRD and GSPD experiments are imposed by prescribing the
155 vegetation type over the Sahara-Sahel region [11°-33°N, 15°W-35°E] to be evergreen shrub
156 with a leaf area index (LAI) equal to 2.6. The change in vegetation cover leads to a decrease in
157 the albedo (averaged into the Sahara-Sahel region) from 0.30 to 0.15 (Table 1). Surface
158 roughness and soil wetness are prescribed to PI values in the sensitivity experiments. As
159 discussed in detail in Pausata et al. [2016], the experiment design is idealized. Nevertheless,
160 recent studies have shown that paleolakes extended at least to 28°N [Lézine et al., 2011] and
161 tropical vegetation migrated as far as 24°N during the AHP, about 1000 km further north than
162 today [Hély et al., 2014]. Based on leaf wax isotopes, Tierney et al. [2017] show that the Green
163 Sahara extended all the way to the northern most part of western Sahara, with peaks in
164 precipitation above 1000 mm/year relative to present day. Hence, proxy evidence suggests an
165 extensive greening across the entire Sahara during the MH, overall supporting the choice of
166 the prescribed vegetation cover in the GSRD and GSPD experiments. Furthermore, Pausata et
167 al. [2016] have shown that changes in WAM strength do not strongly depend on the exact
168 choice of the vegetation type prescribed over eastern Sahara, which may have experienced
169 dryer conditions compared to the western part.

170

171 The atmospheric dust concentration is prescribed in the CTL simulation by using the PI dust
172 climatology derived from the Community Atmosphere Model [Albani et al., 2014], which is
173 used in the CMIP5 exercise. The size distribution of dust corresponds to diameter ranges
174 from 0.1 to 10.0 μm . The model used in this study simulates the direct effect of dust on the
175 atmospheric radiative balance, while the indirect effect on cloud formation and microphysics
176 is not included. To take into account the decreased dust emission and concentration typical of
177 the AHP, in the GSRD and PSRD experiments the PI dust concentration is reduced by 80%
178 throughout the troposphere (up to 150 hPa) over a broad area around the Sahara desert (see
179 Figure S1 in Pausata et al. [2016]). This choice is justified by evidence from proxy records
180 off the coast of Morocco showing a dust flux reduction of 60-80% [deMenocal et al., 2000;
181 McGee et al., 2013], and also supported by a recent dust modelling study [Egerer et al.,
182 2016]. The imposed dust reduction leads to a decrease in the global dust aerosol optical depth
183 (AOD) of almost 60% and in the total AOD of 0.02 (see Figure 1 in Pausata et al. [2016]).
184 The PI climatological dust mixing ratio prescribed in the CTL and GSPD simulations (Figure
185 1a) closely represents the main features of the present-day dust distribution over North
186 Africa, namely the peak over the Bodélé depression north of Lake Chad, and the band over
187 western Sahara and Sahel [Ginoux et al., 2012]. An intense peak in concentration is present
188 over eastern Sahara, largely dominating the African dust emission in summer, which appears
189 as an artifact when compared with satellite retrievals for the present-day distribution
190 (compare Figure 1a with Figure 4 in Ginoux et al. [2012]). This artifact affects the radiative
191 balance simulated in the region, by masking the dust sources to the west, with potential
192 repercussions on the simulation of the regional atmospheric dynamics. Additional simulations
193 have been performed to test the effect of imposing a more realistic dust climatology, finding
194 barely significant differences in the effect of dust reduction on precipitation (see Figures S1-

195 S3 in the Supplementary Material online). However, dust emission from the central and
196 western Sahara is intimately linked to the surface wind pattern over North Africa, which
197 shows a broad variability at multidecadal time scales [Evan et al., 2016]. Thus, substantial
198 differences in the surface wind fields in the PI period could explain this apparently
199 incongruous dust concentration pattern produced by the CAM model. However, the
200 meridional profile of dust concentration (averaged between 15°W and 25°E, to cut the
201 unusual peak to the east off) is close to the present-day pattern, showing a dust layer reaching
202 the mid-troposphere, with an intense peak around 15-20°N (corresponding to the Bodélé and
203 western Sahara sources), and a mid-tropospheric plume extending south to the Gulf of Guinea
204 (Figures 1c, e). When the dust concentration is reduced by 80%, the highest anomalies are
205 located where the PI concentration is maximum, namely over central and eastern Sahara,
206 around 15-20°N, up to 800-700 hPa (Figures 1b, d, f).

207

208 In each experiment, initial conditions are taken from a 700-year PI spin-up run, and the
209 simulations are then run for about 300 years. The quasi-equilibrium is reached after 100-200
210 years depending on the experiment. This study focuses on the monsoonal season (June to
211 September, JJAS) for the last 30 years of each experiment. The experimental setup is
212 summarized in Table 1, and a detailed description is presented in Pausata et al. [2016], along
213 with a more comprehensive discussion of the model's performances.

214

215 **3 Results**

216

217 This section presents the impacts of changes in Saharan vegetation and dust concentration on
218 precipitation and atmospheric dynamics over West Africa in each sensitivity experiment
219 relative to the MH reference simulation (CTL). The responses in the radiative and energy

220 variables are then analyzed. Unless otherwise specified, the significance of the changes is
221 estimated via a two-tailed Student's t-test at the 95% confidence level.

222

223 **3.1 Precipitation and Dynamics**

224

225 In the CTL simulation (Figure 2a), the climatological WAM precipitation is consistent with
226 the “classical” pattern observed in the PI and historical periods [Nicholson, 2013]. The peak
227 precipitation reaches over 16 mm/day on the western coastline around 10°N, and
228 approximately 10 mm/day in the interior, slightly in excess of present-day observations (not
229 shown), but in agreement with the most CMIP5 models [Harrison et al., 2014]. The
230 northernmost extent of the monsoon is just above 16°N, 2° latitude further north than a PI
231 simulation performed using the same model (see Figure S6 in Pausata et al. [2016]). The
232 main change in the precipitation pattern in the GSRD experiment is a dramatic increase of
233 precipitation over the whole North African subcontinent, with significant anomalies spanning
234 10°N to 25°N and extending to the Mediterranean coast in the western sector (Figures 2b, c).
235 The response in the GSPD experiment (Figures 2d, e) is qualitatively similar to the GSRD
236 experiment, though slightly weaker in intensity and extension, suggesting that the dust
237 reduction tends to enhance the vegetation forcing (see the GSRD-GSPD difference in Figure
238 2h). In the Green Sahara experiments, the location of the precipitation maximum is around
239 15°N (Figure 2b, d), consistently with the proxy-based estimation by Hély et al. [2014]. The
240 simulated yearly precipitation anomalies exceed 700 mm around the peak (10°-20°N), and
241 reach 300 mm in the Sahara region (20°-30°N). Skinner and Poulsen [2016] find similar
242 precipitation anomalies extending into the late monsoonal season, pointing out the role of
243 extratropical troughs advecting tropical moisture toward the Sahara in the form of
244 concentrated plumes of water vapor. These values are in line with MH precipitation

245 reconstructions based on diverse paleoclimate proxies, which estimate the difference with the
246 present day over West Africa in the range 300-500 mm [Kröpelin et al., 2008; Harrison et al.,
247 2014; Tierney et al., 2017]. On the other hand, the reduction of dust alone in the PSRD
248 experiment leads to reduced precipitation over the Sahelian belt relative to the CTL case
249 (Figures 2f, g).

250

251 The wind pattern at 925 hPa in the CTL simulation is dominated by the southwesterly moist
252 monsoonal flow from the Tropical Atlantic, the northeasterly Harmattan dry wind, and the
253 cyclonic gyre associated with the Saharan Heat Low (SHL, [Lavaysse et al., 2009]), located
254 around [10°W-10°E, 20°-30°N] (Figure 3a). In the GSRD experiment (Figure 3b), the mean
255 wind pattern displays a general northward shift, relative to the MH CTL simulation, resulting
256 in an intensification of the monsoonal flow and a weakening of the Harmattan wind and the
257 SHL (-2.1 K at the peak). The latter migrates 3.2° to the north (Figure 3c), coherently with
258 the precipitation response shown in Figures 2b, c. In comparison to the GSRD experiment,
259 the GSPD experiment shows a similar response in the wind field (Figure 3h), with a slightly
260 stronger SHL (+0.9 K at the peak), displaced 0.7° to the south (Figure 3d, e). The negative
261 precipitation response in the Sahel simulated in the PSRD experiment (Figures 2f, g) is
262 instead associated with a weakening of the monsoonal flow and a deeper southward
263 penetration of the Harmattan in the eastern Sahel, accompanied by a 0.7° southward
264 migration of the SHL (Figures 3f, g).

265

266 In the CTL simulation, the mid-tropospheric (700 hPa) wind field appears similar, in its main
267 features, to the circulation pattern characterizing the present-day WAM (Figure 4a). The
268 African Easterly Jet (AEJ) axis is located around 16°N, and the anticyclonic gyre of the
269 Saharan high contrasts the SHL in the lower troposphere [Chen, 2005]. The GSRD

270 experiment shows a dramatic reorganization of the circulation, with a significant 6.5°
271 northward shift and 2.7 m/s weakening of the AEJ, and the installation of a westerly flow
272 around 10-15°N. This results in a large-scale cyclonic gyre over West Africa around 15°N,
273 and a substantial weakening of the Saharan high (Figures 4b, c). The wind field response in
274 the GSPD experiment shows a similar structure in comparison with the GSRD experiment,
275 but in this case the Saharan high is slightly stronger and the AEJ is slightly weaker and
276 displaced 1.2° to the south (Figures 4d, e, h). In present-day atmospheric dynamics, the AEJ
277 is maintained by the soil moisture gradient at the surface and the two meridional circulations
278 forced by the dry convection of the SHL to the north and the moist convection associated
279 with the intertropical convergence zone (ITCZ) to the south [Thorncroft and Blackburn 1999;
280 Wu et al. 2009]. Therefore, the weakening and northward shift of the jet in the Green Sahara
281 scenario is the expected response to the vanished surface soil moisture gradient and the
282 substantial reduction of the SHL-associated circulation. In the PSRD experiment, the only
283 significant change is a 1.5° southward shift of the AEJ compared to the CTL (Figures 4f, g).
284 Very similar response patterns are observed at 500 hPa (Figure S4), pointing to the
285 emergence of an intense westerly flow around 10°N from low levels up to the mid-
286 troposphere, and a remarkable limitation in the southward extension of the Saharan high in
287 the GSRD and GSPD experiments.

288

289 The upper tropospheric circulation at 200 hPa is dominated by the Tropical Easterly Jet (TEJ)
290 axis between 0 and 15°N over Africa, the subtropical jet from the Mediterranean Sea to mid-
291 latitudes in western Asia, and the subpolar jet at mid-latitudes over the North Atlantic (Figure
292 5a). The GSRD and GSPD experiments (Figures 5b-e) show a general intensification of the
293 zonal circulation in the upper troposphere, characterized by a northward shift of the
294 subtropical jet and the emergence of an easterly jet streak around 15°N over West Africa. The

295 responses in GSRD and GSPD are not significantly different (Figures 5c, e, h). The response
296 of the zonal circulation in the upper troposphere to the greening of the Sahara is interpreted as
297 the upper branch of an anomalous Walker-like circulation (Figure S5), generated by the
298 widespread intensification of deep convection, and connected with the strengthened westerly
299 flow in the lower troposphere (see Figures 3c, e). In the PSRD experiment, the zonal
300 circulation is weakened and shifted slightly southward (Figures 5f, g).

301

302 The dramatic reorganization of the regional circulation in the GSRD experiment is also
303 revealed by the meridional overturning circulation averaged over West Africa [10°W, 10°E]
304 (Figure 6). The CTL experiment simulates a circulation pattern resembling the large scale
305 dynamics associated with the present-day WAM (Figure 6a): (i) the southern hemisphere
306 Hadley cell, (ii) the southerly monsoonal flow in the lower troposphere reaching the Sahel,
307 (iii) the deep convection associated with the ITCZ, (iv) the dry shallow convection over the
308 Sahara associated with the SHL, contrasting with the subsidence associated with the Saharan
309 high in the mid-troposphere (500-600 hPa) [Nicholson, 2013]. In the GSRD and GSPD
310 experiments, the change in surface conditions over the Sahara leads to an extension of the
311 deep convection up to 30-35°N, which cancels the mid-tropospheric divergence associated
312 with the SHL shallow convection contrasting the Saharan high subsidence (Figures 6b, d).
313 The reduction in dust concentration amplifies the response to the land cover change, inducing
314 stronger anomalies in deep convection over the Sahel and the Sahara desert (Figures 6c, e, h).
315 The PSRD experiment does not show any substantial systematic changes in the atmospheric
316 meridional overturning circulation compared to the CTL experiment (Figure 6f, g).

317

318 The analysis discussed above shows that a greening of the Sahara has a dramatic impact on
319 West African precipitation, introducing unusual features in the monsoon-related regional

320 circulation. Summarizing, rainfall intensifies over the Sahara and extends all the way up to
321 the Mediterranean coast (Figure 2), and some of the dynamical features characterizing the
322 “classical” WAM picture are substantially altered. The monsoonal flow extends far to the
323 north and into the mid-troposphere, weakening the AEJ and shifting it northward by 6.5°
324 (Figures 3 and 4). Moreover, the mid-tropospheric contrast between the SHL convection and
325 the Saharan high subsidence vanishes (Figures 3 and 4). These changes appear as a general
326 reorganization of the circulation towards a cyclonic gyre throughout the troposphere,
327 resulting from the development of deep convection over the Sahara (Figure 6). Patricola and
328 Cook [2007] described similar modifications to the monsoonal dynamics during the MH, by
329 prescribing shrubland north of approximately 20° in regional climate model simulations. The
330 present study highlights that the reduction in dust concentration, following the greening of the
331 Sahara, amplifies the impact of land cover change. On the other hand, the effect of reduction
332 in dust alone results in a weak decrease in the Sahelian precipitation, accompanied by small
333 changes in the atmospheric circulation. This suggests a peculiar role of dust reduction in
334 amplifying the anomalies when associated with land cover changes.

335

336 The changes in the WAM atmospheric dynamics are accompanied by important changes in
337 the mechanisms driving the precipitation. A major source of synoptic-scale rainfall variability
338 throughout present-day West Africa is associated with the African easterly waves (AEWs):
339 westward propagating disturbances with wavelengths of approximately 3000 km and periods
340 below 5-6 days [Diedhiou et al., 1998]. A longer-period regime of 6-9 day waves has also
341 been identified in the region [Diedhiou et al., 1998; 1999], but these typically display a
342 weaker association with precipitation [Gu et al., 2004]. The AEWs are particularly important
343 over the Sahel, where they modulate rainfall through the initiation and organization of
344 mesoscale convective systems (MCSs) and squall lines during the monsoonal season [Skinner

345 and Diffenbaugh, 2013]. AEW activity is detected in each experiment by computing the
346 variance of the daily meridional wind at 700 hPa, after the application of a 2.5-5 day
347 bandpass filter. As in Diedhiou et al. [1998, 1999] and Gu et al. [2004], “wavy” days are
348 defined here as the days where the filtered southerly wind is at maximum and exceeds 0.5
349 m/s at a reference point. Such point is chosen to be representative of the main track of wave
350 activity as defined by the filtered meridional wind in the different simulations. For the CTL
351 and PSRD cases, the location [15°N, 0°E] is selected, similarly to studies analyzing the
352 present-day monsoon (e.g. Diedhiou et al., [1999]). For the GSPD and GSRD simulations,
353 the location [20°N, 0°E] is selected, reflecting the significant northward shift of the
354 monsoonal circulation. It is noted that the results are relatively insensitive to small meridional
355 shifts in the reference point (not shown, see also Gu et al., [2004]). The precipitation is then
356 averaged over the selected days and the JJAS climatology for each experiment is subtracted,
357 in order to calculate the precipitation anomalies associated with wavy days in mm/day. As a
358 caveat, it is noted that this methodology implicitly assumes that all the precipitation occurring
359 during a wavy day can be ascribed to MCSs associated to AEWs.

360

361 The CTL simulation displays a band of wave activity that originates west of the Ethiopian
362 highlands and maximizes over western Sahel, propagating across the Tropical Atlantic Ocean
363 (Figure 7a). This is similar to what seen in reanalysis data, albeit with larger variance values
364 (see Figure 3 in Skinner and Diffenbaugh [2013]). Both the GSRD and GSPD experiments
365 (Figures 7b-e) show a northward shift and a marked weakening of the meridional wind
366 variance in the region affected by the AEW activity. The difference between GSRD and
367 GSPD indicates that the dust reduction favors a further weakening of the AEW activity in the
368 presence of a vegetated Sahara (Figure 7h). Conversely, the PSRD experiment presents a
369 weakened AEW pattern relative to the CTL simulation (Figures 7e, f).

370

371 Since the definition of a wavy day focuses on meridional wind maxima, their frequency is
372 similar across all simulations, ranging from 25.5% in the CTL simulation to 24.0% in the
373 PSRD case. However, the intensity of these AEW episodes shows a broader spread, with the
374 CTL simulation displaying a mean filtered meridional wind velocity of 2.6 m/s compared to
375 the 2.3 m/s of the GSPD case and the 2.1 m/s of the GSRD case. A one-sided two-sampled
376 Kolmogorov-Smirnov test confirms that the cumulative distribution function of the AEW
377 intensity for the GSPD case is larger than that of the CTL at the 95% confidence level,
378 indicating a shift of the distribution towards lower values.

379

380 The significant differences in the filtered meridional wind variance and AEW intensity result
381 in equally large changes in the precipitation associated with the AEWs. The precipitation on
382 wavy days is characterized by a multipolar anomaly pattern, clearly showing the signature of
383 a propagating wave (Figure 8). The peak precipitation anomalies driven by the AEW-
384 associated MCSs are not co-located with the reference points at which the waves are
385 identified. Rather, the peaks tend to occur to the west or south-west of the reference points, in
386 close agreement with previous studies (e.g., Diedhiou et al. [1999] and Gu et al. [2004]). The
387 peak average precipitation anomaly in the CTL simulation is 4.7 mm/day (Figure 8a), and
388 wavy days account for 41.5% of the total seasonal precipitation at that location (Figure 8b).
389 The GSPD and GSRD simulations display a clear northward shift of the dipole, with a sizable
390 reduction of the average peak precipitation anomalies on wavy days to 1.6 and 2.3 mm/day,
391 respectively (Figures 8c, e). These correspond to 36.5% and 33.3% of the total precipitation
392 (Figures 8d, f), respectively. The dust reduction alone leads to slight changes in this pattern,
393 with a 4.1 mm/day peak anomaly, representing 39.2% of the total precipitation.

394

395 The decreased importance of AEWs under Green Sahara conditions indicates a shift towards
396 variability on time scales longer than the synoptic one. This is confirmed by a spectral
397 analysis of precipitation and meridional wind at 700 hPa over [5°-16°N, 10°W-10°E] for the
398 CTL and PSRD simulations and [5°-28°N, 10°W-10°E] for the GSRD and GSPD
399 simulations. The different domains, as for the wavy day calculation, reflect the northward
400 shift of the monsoon under a Green Sahara. The northern limits are chosen to roughly match
401 the monsoon's northernmost extent, calculated following the methodology of Pausata et al.
402 [2016]. The CTL and PSRD meridional wind spectra display a clear peak around the AEW
403 periodicity (2-5 days), while in the two Green Sahara scenarios the power in this band is
404 weaker, and the spectra increase almost monotonically with period (Figure 9b). The
405 precipitation spectra (Figure 9a) reflect this difference. In the CTL and PSRD simulations
406 there is again a clear peak matching the AEW periodicity and highlighting the waves'
407 dominant role in driving precipitation. In the two Green Sahara scenarios, the spectra show a
408 roughly monotonic increase with period, suggesting that the 2-5 day band only accounts for a
409 modest fraction of the total precipitation. . Furthermore, while longer waves in the 6-9 day
410 band grow in importance under a green Sahara, they do not play the same role as the 2-5 day
411 band in the desert Sahara scenarios (Figure 9). The Green Sahara simulations display a higher
412 spectral power than the desert Sahara in both meridional wind and precipitation beyond ~6
413 days, but there is no clear peak to suggest that a specific dynamical mechanism is acting to
414 enhance the precipitation over this frequency range.

415

416 Summarizing, the Green Sahara scenario is characterized by a widespread increase in
417 precipitation, accompanied by a significant reduction in the 3-5 day AEW activity and in
418 their ability to drive precipitation. The precipitation and meridional wind spectra show a shift
419 towards longer periods and do not display any prominent peaks. This is somewhat surprising,

420 because a sizable fraction of the present-day monsoonal precipitation is ascribed to the
421 convection triggered by AEWs [Skinner and Diffenbaugh, 2013]. Hence, a more active AEW
422 regime might be expected under an intensified MH monsoon, due to the increased latent heat
423 and convection associated with the increased precipitation. This apparent contradiction can be
424 reconciled by considering the role of the AEJ in triggering and strengthening the AEWs,
425 through barotropic-baroclinic instabilities [Thorncroft and Hoskins, 1994]. Wu et al. [2012]
426 found that synoptic disturbances are more likely where the zonal flow is stronger and that the
427 more unstable waves are likely to be closer to the zonal flow maximum, i.e., closer to an
428 intense AEJ core. It is therefore argued that, in the Green Sahara scenarios, the radical
429 modification of the AEJ dynamics dominates the thermodynamic changes; the sizable
430 northward migration (+6.5°) and weakening (-2.7 m/s) being more effective in shifting and
431 suppressing the wave activity than the latent heat and convection increase in enhancing it.
432 Finally, it is highlighted that the dust reduction induces opposite responses in the atmospheric
433 dynamics, depending on the land cover of the Sahara: it leads to more or less favorable
434 conditions for precipitation, when associated with a vegetated or desert surface, respectively.

435

436 **3.2 Energy budget**

437

438 The rainfall and atmospheric circulation patterns simulated in the sensitivity experiments are
439 closely linked to the changes in surface conditions and radiative properties of the atmosphere,
440 which in turn affect the energetics and dynamics of the WAM. This section elucidates the
441 links between energetics, precipitation and dynamics in the sensitivity experiments, through
442 the analysis of the atmospheric radiative and heat fluxes, and moist static energy (MSE)
443 content.

444

445 The greening of the Sahara (GSPD simulation) lowers the surface albedo over a vast portion
446 of North Africa (see Table 1), leading to an increase in the net flux of the clear-sky
447 downward SW radiation at the top of the atmosphere (TOA) (Figure 10c). The downward
448 SW flux anomaly is almost uniform and limited to the area where the surface albedo is
449 lowered. The reduction in dust concentration, when the Sahara is deserted (PSRD
450 simulation), leads to an upward SW flux anomaly at the TOA, in which the fingerprint of the
451 dust reduction pattern is evident (Figure 10e). The upward SW flux is due to the increase in
452 the planetary albedo produced by the removal of airborne dust over a bright surface. On the
453 contrary, the removal of airborne dust when the Sahara is vegetated (GSRD simulation)
454 further reduces the planetary albedo of a dark surface, enhancing the downward flux anomaly
455 of SW radiation simulated in the GSPD experiment (Figure 10a), although the effect is quite
456 weak (Figure 10g).

457

458 At the surface, the effect of greening the Sahara (GSPD simulation) is similar to the TOA,
459 resulting in a positive anomaly of the downward SW radiation, limited to the area where the
460 surface albedo is reduced (Figure 10d). A widespread increase in the downward SW radiation
461 at the surface is also simulated in the PSRD experiment (Figure 10f), in which the removal of
462 airborne dust reduces the absorption of the solar radiation, especially at latitudes around
463 15°N, where dust concentration is maximum (see Figure 1). By reducing the absorption of
464 solar radiation, the removal of airborne dust over the Green Sahara (GSRD simulation)
465 enhances the positive anomaly in the downward SW radiation due to the lower surface albedo
466 (Figure 10b, h).

467

468 In the Green Sahara experiments, the SW radiative forcing simulated at the surface over the
469 Sahel and Sahara is accompanied by a negative (i.e. upward) anomaly in the sensible heat

470 (SH) flux, which reverts to positive (i.e. downward) between 10°N and 20°N due to the
471 cooling effect of the precipitation anomalies (Figures 11a, c; cf. Figure 2). The additional SW
472 radiative forcing in the GSRD simulation, associated with the dust reduction, limits the
473 upward SH flux in the eastern Sahel (Figures 11a, g). In the PSRD simulation, the upward SH
474 flux anomaly is widespread, peaking between 10°N and 20°N, where the SW radiative
475 forcing is maximum (Figure 11e; cf. Figure 10f). The latent heat (LH) flux at the surface
476 shows significant negative (i.e. upward) anomalies in the Green Sahara scenarios (Figures
477 11b, d). The release of LH into the atmosphere is associated with enhanced evaporation over
478 the vegetated surface, induced by the SW radiative forcing. The LH response is shaped by the
479 precipitation anomalies, further enhancing evaporation (Figures 11b, d; cf. Figure 2). The
480 reduction in dust concentration over the Green Sahara increases SW radiative forcing and
481 precipitation at the surface, strengthening the LH release into the atmosphere (Figure 11h).

482 Reduction in dust concentration alone (PSRD simulation) results in LH release over the
483 vegetated Sahel (south of 15°N), induced by the SW radiative forcing, while the decreased
484 precipitation leads to negative anomalies in the LH flux north of 15°N (Figure 11f).

485

486 The combination of the LH and SH fluxes results in an intense upward heat flux over the
487 Sahara in the Green Sahara experiments, and a weaker upward flux peaking over the Sahel in
488 the PSRD experiment (Figure S6). The heat flux then triggers moist convection, whose
489 modifications are diagnosed in the sensitivity experiments through the MSE content of the
490 atmosphere. This is defined as the sum of geopotential, enthalpy and latent heat:

491
$$MSE = gz + C_p T + Lq,$$

492 where g is the gravitational acceleration, z the geopotential height, C_p the specific heat of dry
493 air at constant pressure, T temperature, L the latent heat of evaporation, and q the specific
494 humidity. The MSE is a direct indicator of the monsoonal precipitation, because the

495 transformation of enthalpy and latent energy available in the lower troposphere into
496 geopotential energy aloft is the main signal of convection [Fontaine and Philippon, 2000]. In
497 the GSRD simulation, the meridional profile of the MSE atmospheric content shows an
498 intense positive anomaly dominating the entire troposphere from 15°N to 35°N (Figure 12a).
499 Close to the surface, at 1000 hPa, the MSE peak moves 6° northward (from 14°N in CTL to
500 20°N; Figure S7), accompanied by an increase in the meridional difference between the peak
501 and the Guinea Coast (+1.7 kJ/kg; Figure S7). The GSPD simulation shows a similar pattern,
502 characterized by weaker anomalies (Figures 12 b, d): a +5° migration of the MSE peak and a
503 1.2 kJ/kg increase in the meridional difference with the Guinea Coast (Figure S7). These
504 modifications elucidate the precipitation response in the Green Sahara experiments: the
505 meridional migration and intensification of the peak MSE favor the northward penetration of
506 the precipitation belt, and the MSE excess in the lower-middle troposphere triggers deep
507 convection over the Sahara. In the PSRD simulation (Figure 12c), the MSE content shows a
508 negative anomaly in the middle-upper troposphere, and a weakening of the maximum close to
509 the surface (-0.7 kJ/kg at 1000 hPa; Figure S7), resulting in a weaker monsoonal circulation.

510

511 **4 Discussion and conclusions**

512

513 The objective of this study is to improve the understanding of the changes in the West
514 African monsoon dynamics and precipitation patterns during the mid-Holocene, when large
515 part of North Africa was characterized by a mesic environment and the Sahara desert was
516 extensively covered by vegetation. A set of idealized sensitivity experiments, performed with
517 the EC-Earth climate model, has been used to understand the changes in atmospheric
518 circulation that characterized that period. In these simulations, vegetation cover over the

519 Sahara and a reduction in atmospheric mineral dust concentrations are imposed together and
520 in turn.

521

522 By using these set of simulations, Pausata et al. [2016] showed that the vegetation cover over
523 the Sahara is the primary driver of a radical alteration of the West African monsoon during
524 the mid-Holocene, with the associated dust reduction further strengthening precipitation
525 anomalies inland. The precipitation simulated in their “Green Sahara and reduced dust”
526 experiment is indeed quantitatively consistent with estimations of the mid-Holocene
527 precipitation based on paleoclimate proxy data [Tierney et al., 2017]. This study further
528 extends the analysis performed by Pausata et al. [2016], to improve the understanding of the
529 changes in the monsoonal dynamics accompanying the precipitation anomalies. The
530 modifications of the radiative balance at the surface and the TOA are also discussed, along
531 with the consequent response of the energetics of the monsoonal system.

532

533 When a vegetation cover is prescribed over the Sahara (GSPD), a dramatic response in the
534 West African monsoon is simulated. The lowering of the surface albedo induces an intense
535 positive radiative forcing at the TOA and at the surface over the Sahara, leading to a release
536 of sensible and latent heat into the atmosphere. The heat flux anomaly triggers moist
537 convection over this region, driving a northward expansion of the monsoonal precipitation
538 belt, which extends up to the Mediterranean coast. A substantial modification of the regional
539 circulation is also simulated, with critical alterations in some of the “classical” dynamical
540 features of the West African monsoon, such as the Saharan heat low, the African easterly jet,
541 and the African easterly wave regime. In particular, a weakening of the African easterly wave
542 activity is associated with a decrease in wave-related precipitation, suggesting a reduction of
543 the African easterly waves’ role in triggering and organizing mesoscale convective systems.

544 These changes follow a general reorganization of the circulation towards a cyclonic gyre
545 throughout the troposphere. This analysis suggests that a vegetated Sahara may radically
546 change the regional circulation pattern over West Africa.

547

548 On the opposite, dust reduction alone (PSRD) leads to a slight weakening in the monsoonal
549 circulation and precipitation rates. This weakening is associated with a positive surface
550 radiative forcing in the Sahel and a radiative cooling in the middle-upper troposphere over the
551 Sahara, resulting in a reduction of the meridional energy gradient over West Africa, which
552 prevent monsoonal precipitation to penetrate inland.

553

554 The effect of dust removal on the radiative balance of the atmosphere depends on the
555 radiative properties of the airborne dust in association with the underlying land cover. Indeed,
556 in PI conditions dust reduction increases the planetary albedo above the Sahara desert,
557 cooling the middle-upper troposphere, and induces a positive radiative forcing at the surface,
558 peaking in the vegetated Sahel. On the contrary, when a vegetated surface is prescribed from
559 the Sahel to the Sahara, dust reduction lowers the planetary albedo, warming the middle-
560 upper troposphere, and induces a positive radiative forcing at the surface, peaking in the
561 Sahel-Sahara transition zone where PI dust concentration is maximum.

562

563 The radical modification of the atmospheric circulation over North Africa discussed in this
564 study is likely to have triggered a chain of global-scale changes during the mid-Holocene. For
565 example, Muschitiello et al. [2015], using the same model as in this study, show that the
566 desertification of the Sahara at the end of the African humid period led to a reduction in the
567 atmospheric and oceanic poleward heat transport from tropical to high northern latitudes. The
568 reduction in the poleward heat transport weakened the mid-latitude westerlies, resulting in

569 more negative Arctic Oscillation-like conditions and a consequent cooling over the Arctic
570 and North Atlantic regions. Results presented in this paper show that the greening of the
571 Sahara triggers a strong reduction in African easterly wave activity, which is known to be
572 related with the formation of tropical cyclones in the Atlantic Ocean [Price et al., 2015].
573 Therefore, the decreased AEW activity may induce a reduction in tropical storms developing
574 in the North Atlantic. On the other hand, the reduction in the Saharan dust emission and
575 transport across the tropical Atlantic is favorable to the development of tropical cyclones in
576 the basin, through the radiative warming of the ocean surface [Evan et al., 2006], indicating
577 possible competing effects of the vegetation-dust changes in the Sahara on the Atlantic
578 hurricane season. Furthermore, reducing African easterly waves may initially lead to fewer
579 tropical cyclones but, because of that, the ocean would warm, compensating for the reduced
580 wave activity (assuming no other changes in the thermodynamic structure of the atmosphere).
581 Therefore, the combination of Sahara greening and reduced dust emission during the mid-
582 Holocene may have led to an increased number of tropical cyclones in the Atlantic. By
583 analyzing the same set of simulations used in this study, Pausata et al. [2017a] show indeed
584 an increase in the tropical cyclone activity in both the hemispheres and particularly over the
585 Caribbean basin and the east coast of North America. This hypothesis cannot be tested with
586 current paleotempestology archives that are geographically and temporally patchy, and hence
587 unable to provide a complete view of the Holocene storm activity. However, a proxy-data
588 reconstruction from Puerto Rico suggests a positive correlation between West African
589 monsoon strength and tropical cyclone intensity over the last 5500 years [Donnelly &
590 Woodruff, 2008]. In particular, the period between 2500 and 1000 years BP was the most
591 active tropical storm interval in Puerto Rico and was characterized by a stronger WAM and a
592 reduced Saharan dust flux relative to the following millennium [Mulitza et al., 2010]. In
593 addition, Williams et al. [2016] emphasized the relationship between reduced dust loading

594 during the early and mid-Holocene and the northward expansion of the ITCZ, which in turn
595 can intensify tropical storm activity in the western North Atlantic [van Hengstum et al.,
596 2016]. Pausata et al. [2017b] also show reduced El Niño-Southern Oscillation (ENSO)
597 variability and La Niña-like conditions in winter in a Green Sahara scenario, compared to
598 pre-industrial and the standard CMIP/PMIP mid-Holocene scenarios. These changes in
599 ENSO mean state and variability are in general favorable to tropical storm activity in the
600 North Atlantic basin [Chiacchio et al., 2017]. Further efforts both from the modelling and the
601 proxy community are necessary to better understand the remote teleconnections associated to
602 the West African monsoon variability.

603

604 The findings of this study are also useful also in a future perspective, at regional and global
605 scales. Indeed, the possible positive trend in Sahelian precipitation projected for the 21st
606 century [Biasutti, 2013] may be accompanied and amplified by land cover and dust reduction
607 feedbacks, which in turn may impact climate in remote regions.

608

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610

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624

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626

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817

818 **Tables**

819

820 Table 1: Experimental design used in the control simulation (CTL) and the sensitivity
 821 experiments: present-day Sahara and reduced dust (PSRD); Green Sahara and reduced dust
 822 (GSRD); Green Sahara and pre-industrial dust (GSPD).

Simulation	Orbital forcing (years BP)	GHGs	Vegetation type	Albedo	LAI	Saharan dust
CTL	6k	MH	Desert	0.30	0	PI
PSRD	6k	MH	Desert	0.30	0	80% reduced
GSRD	6k	MH	Evergreen shrub	0.15	2.6	80% reduced
GSPD	6k	MH	Evergreen shrub	0.15	2.6	PI

823

824 **Figures**

825

826 Figure 1: (left) JJAS pre-industrial dust mixing ratio [kg/kg], prescribed in the CTL and
827 GSPD simulations, and (right) difference between reduced dust and pre-industrial
828 concentration. (a, b) Column integral, (c, d) vertical profile averaged in the range [15°W,
829 25°E], (e, f) meridional profile of the column integral.

830

831 Figure 2: JJAS precipitation [mm/day]: (left) climatological means for the CTL simulation
832 and the sensitivity experiments, and (right) differences between the sensitivity experiments
833 and the CTL simulation. In panel (h), the difference between the GSRD and GSPD
834 experiments is shown. Significance at 95% confidence level is displayed.

835

836 Figure 3: JJAS wind field at 925 hPa [m/s]: (left) climatological means for the CTL
837 simulation and the sensitivity experiments, and (right) differences between the sensitivity
838 experiments and the CTL simulation. In panel (h), the difference between the GSRD and
839 GSPD experiments is shown. Significance at 95% confidence level is displayed. Shadings
840 indicate the Saharan heat low location and intensity, represented through the values of air
841 temperature at 850 hPa [K] exceeding the 90th percentile in the domain [0-40°N, 20°W-
842 20°E] [Lavaysse et al., 2009]. The SHL position and intensity is determined by detecting the
843 maximum of the air temperature at 850 hPa inside the domain.

844

845 Figure 4: JJAS wind field at 700 hPa [m/s]: (left) climatological means for the CTL
846 simulation and the sensitivity experiments, and (right) differences between the sensitivity
847 experiments and the CTL simulation. In panel (h), the difference between the GSRD and
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849 indicate the Saharan high location and intensity, represented through the geopotential height
850 at 700 hPa exceeding 3150 m. Red lines represent the mean positions of the African easterly
851 jet in the CTL simulation (dashed) and the sensitivity experiments (solid). The mean axis of
852 the African easterly jet is determined in the domain [20°W-20°E, 0°-40°N] by detecting, at
853 each longitude, the latitudinal location of the minimum of the zonal wind at 700 hPa, and
854 averaging along the zonal direction.

855

856 Figure 5: JJAS wind field at 200 hPa [m/s]: (left) climatological means for the CTL
857 simulation and the sensitivity experiments, and (right) differences between the sensitivity
858 experiments and the CTL simulation. In panel (h), the difference between the GSRD and
859 GSPD experiments is shown. Significance at 95% confidence level is displayed.

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861 Figure 6: JJAS meridional overturning circulation, represented through the meridional wind
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864 differences between the sensitivity experiments and the CTL simulation. In panel (h), the
865 difference between the GSRD and GSPD experiments is shown. Significance at 95%
866 confidence level is displayed.

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868 Figure 7: JJAS African easterly wave activity, represented through the variance of the
869 meridional wind at 700 hPa filtered in the 2.5-5-day band [m²/s²]: (left) climatological means
870 for the CTL simulation and the sensitivity experiments, and (right) differences between the
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879 crosses. Only values in a 15° latitudinal band around the maximum of the meridional wind
880 variance (see Figure 7) are displayed.

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882 Figure 9: Fast Fourier transform analysis of grid-point JJAS daily (a) precipitation and (b)
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884 simulations and [5°-28°N, 10°W-10°E] for the GSRD and GSPD simulations.

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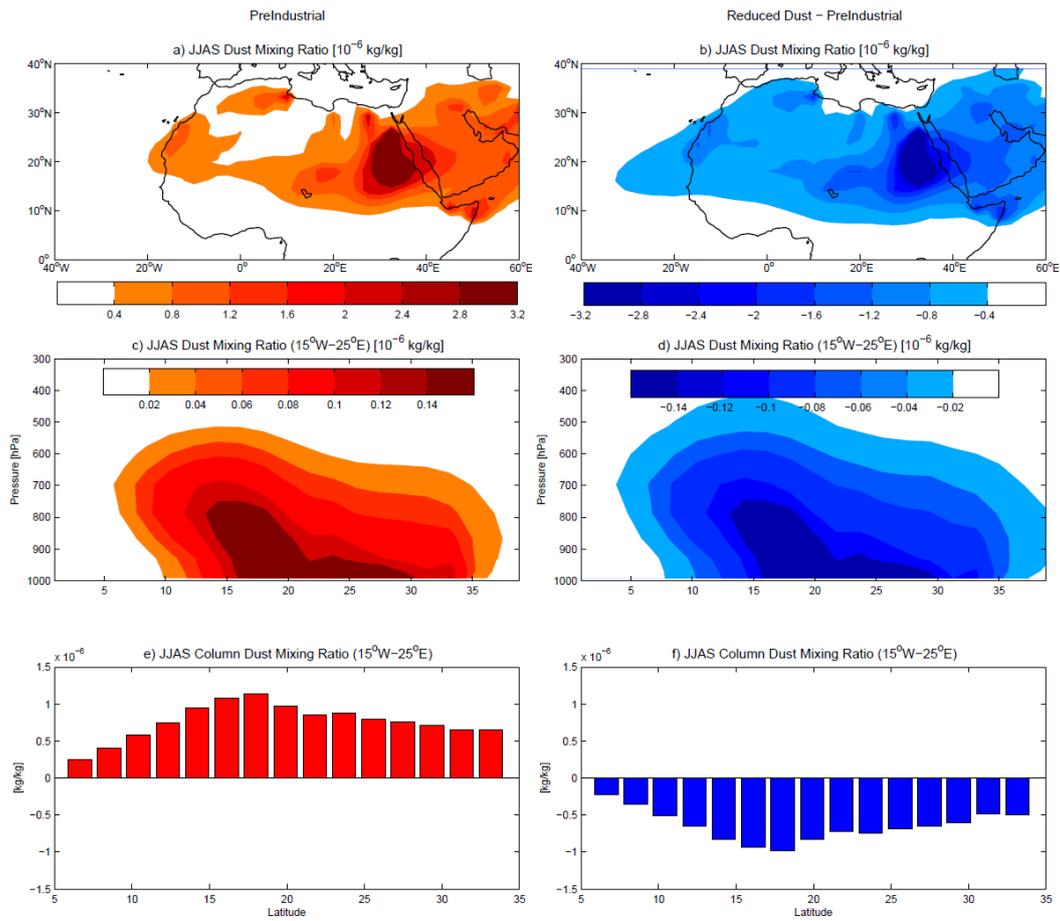
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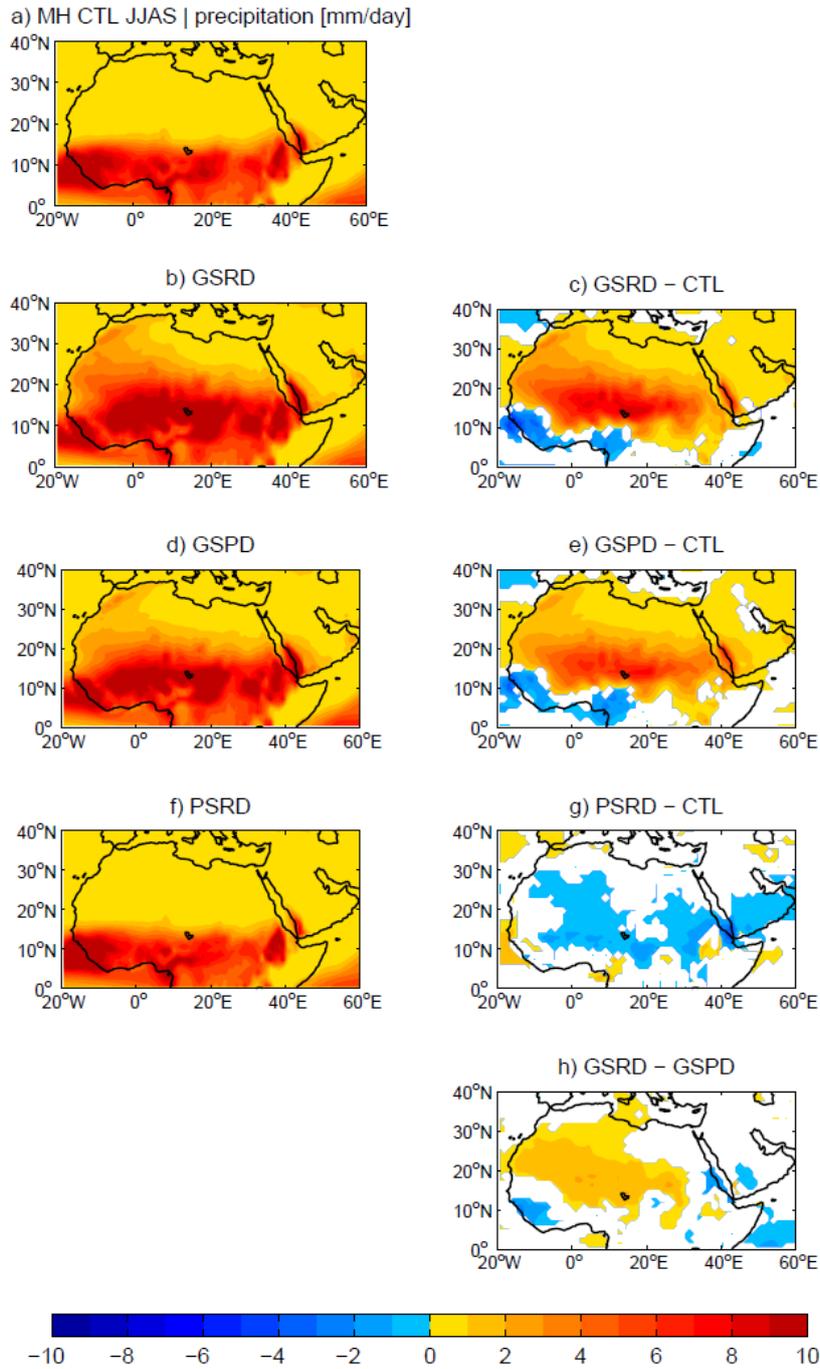
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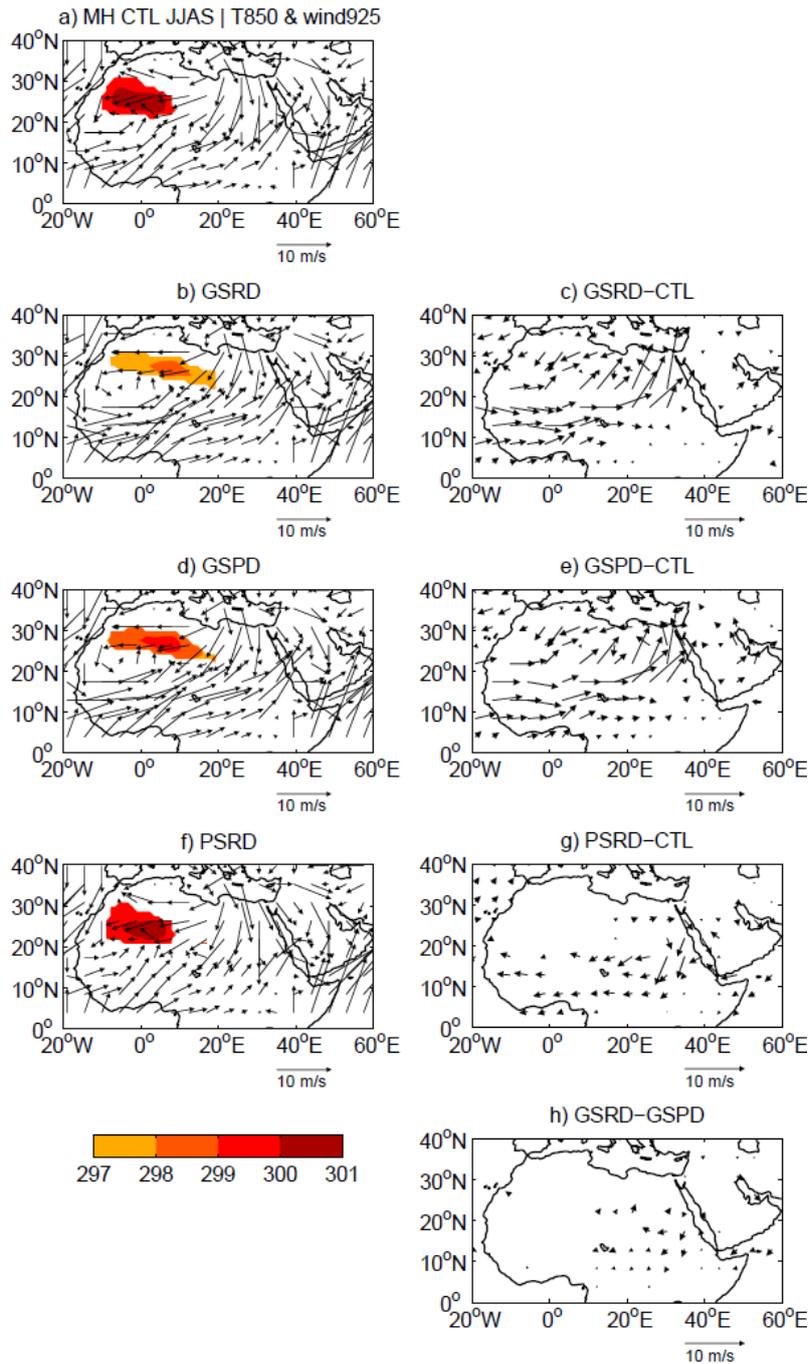
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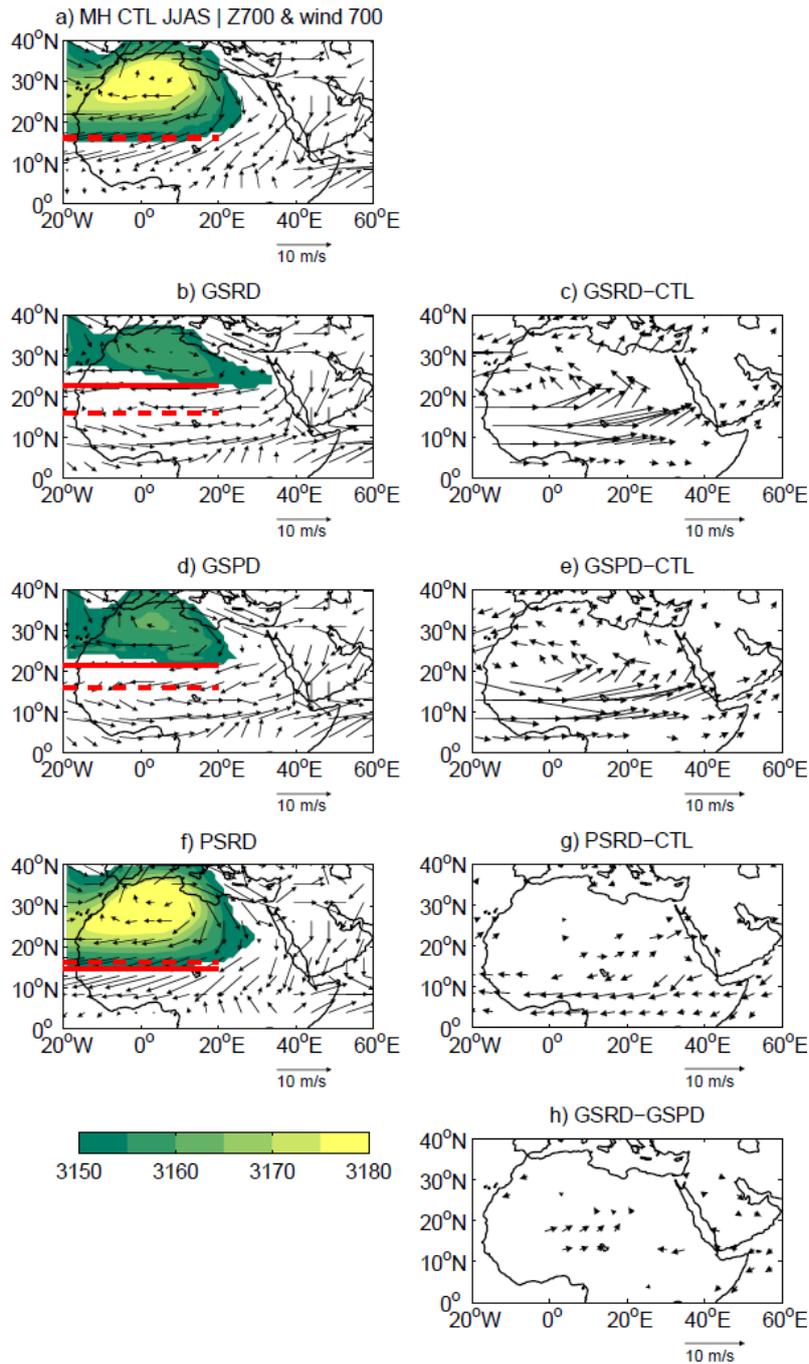
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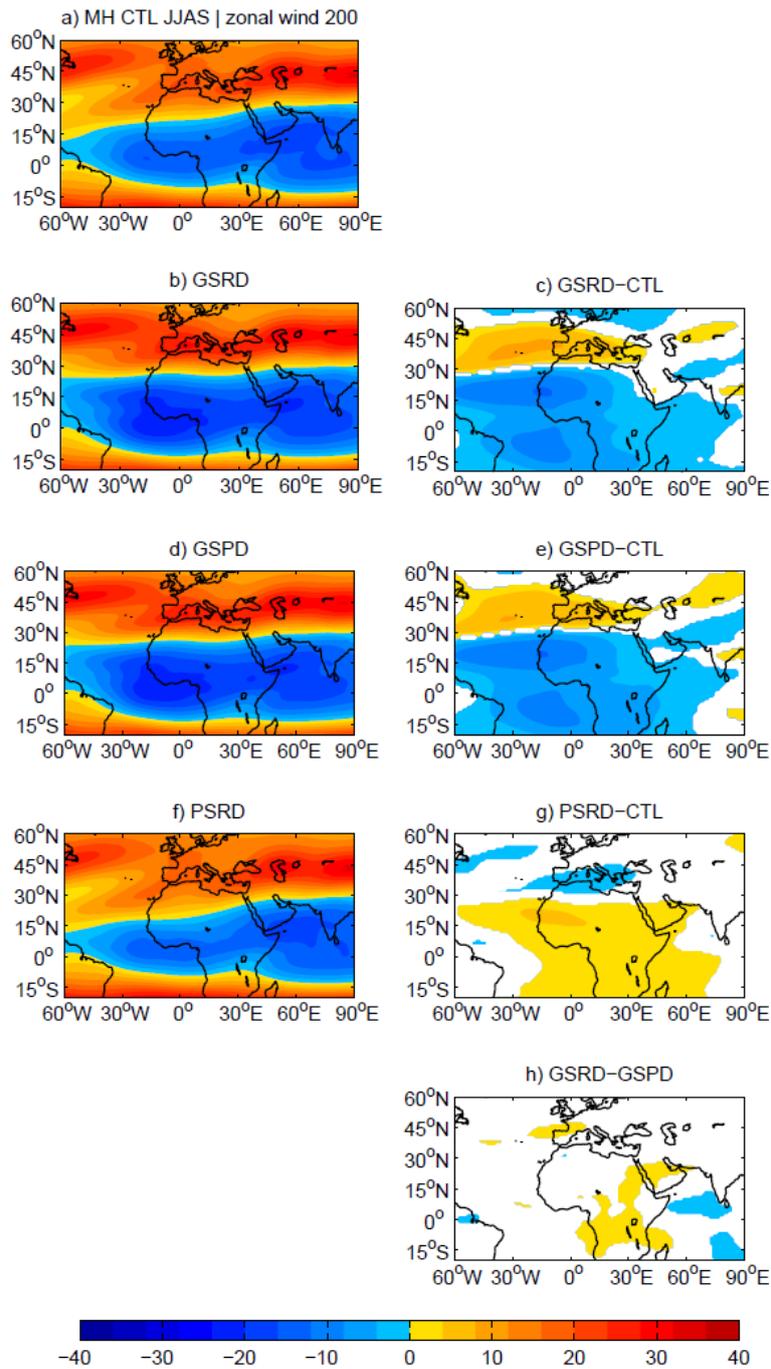
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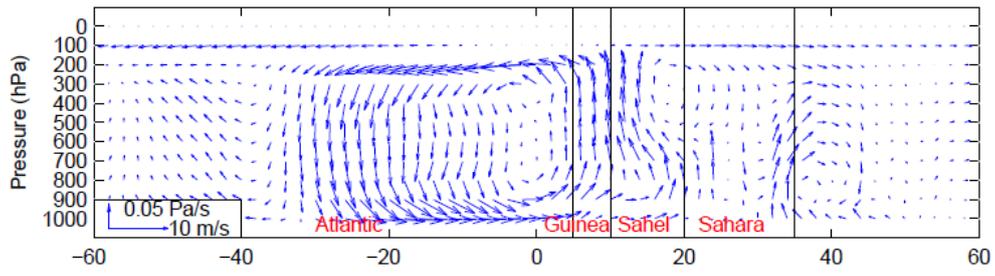


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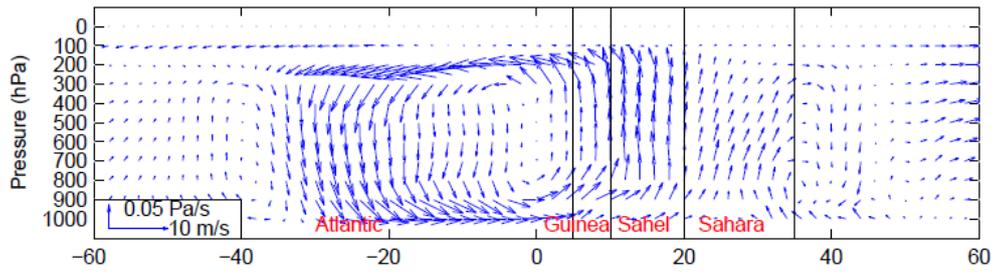
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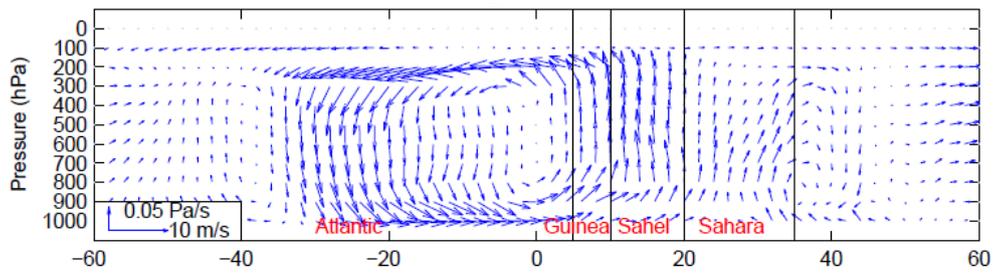
a) MH JJAS CTL ($v, -\omega$) $10^{\circ}\text{W}-10^{\circ}\text{E}$



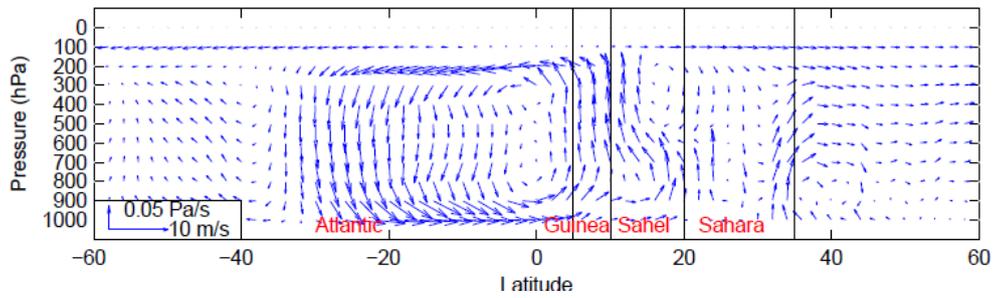
b) GSRD

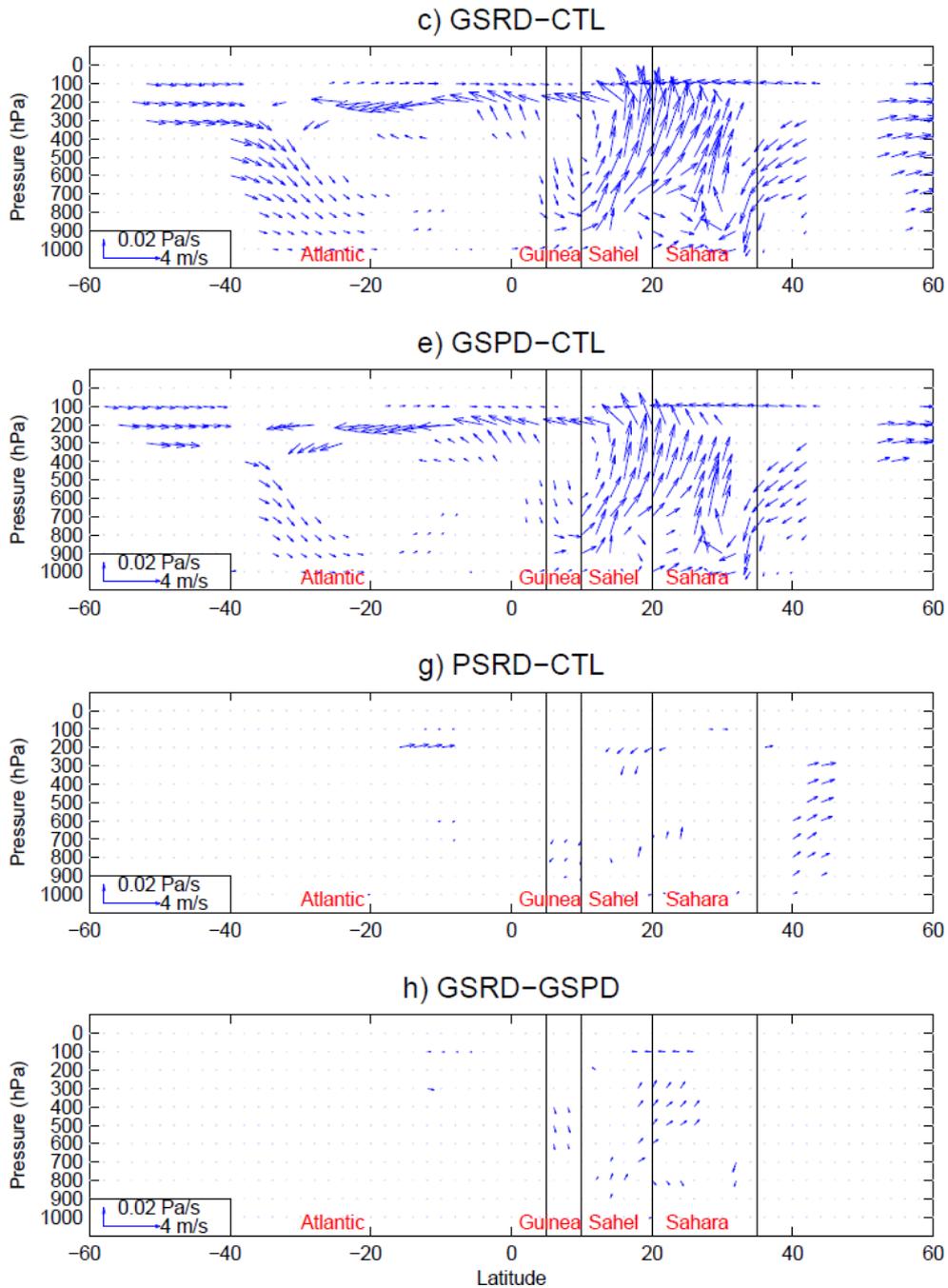


d) GSPD



f) PSRD





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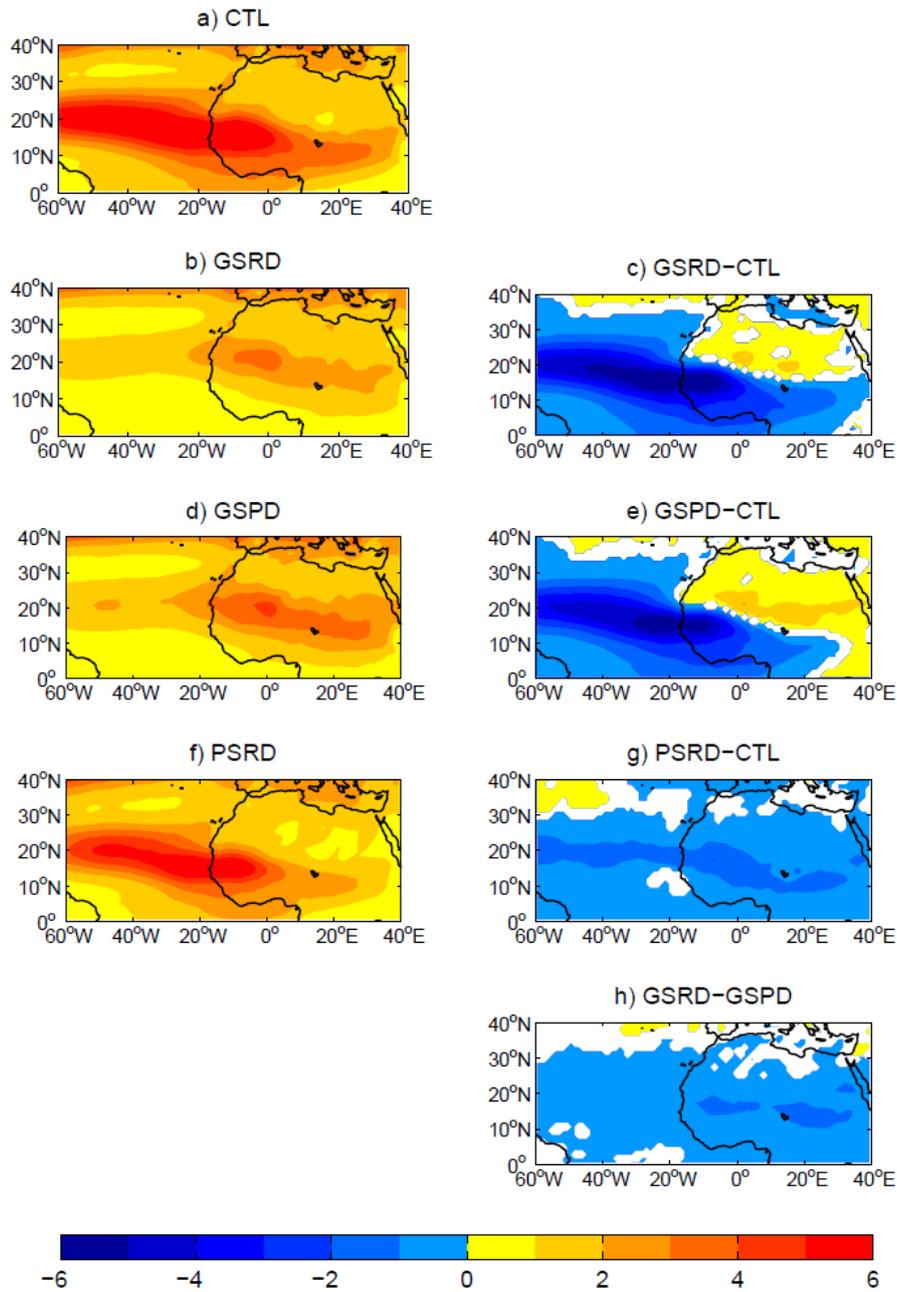
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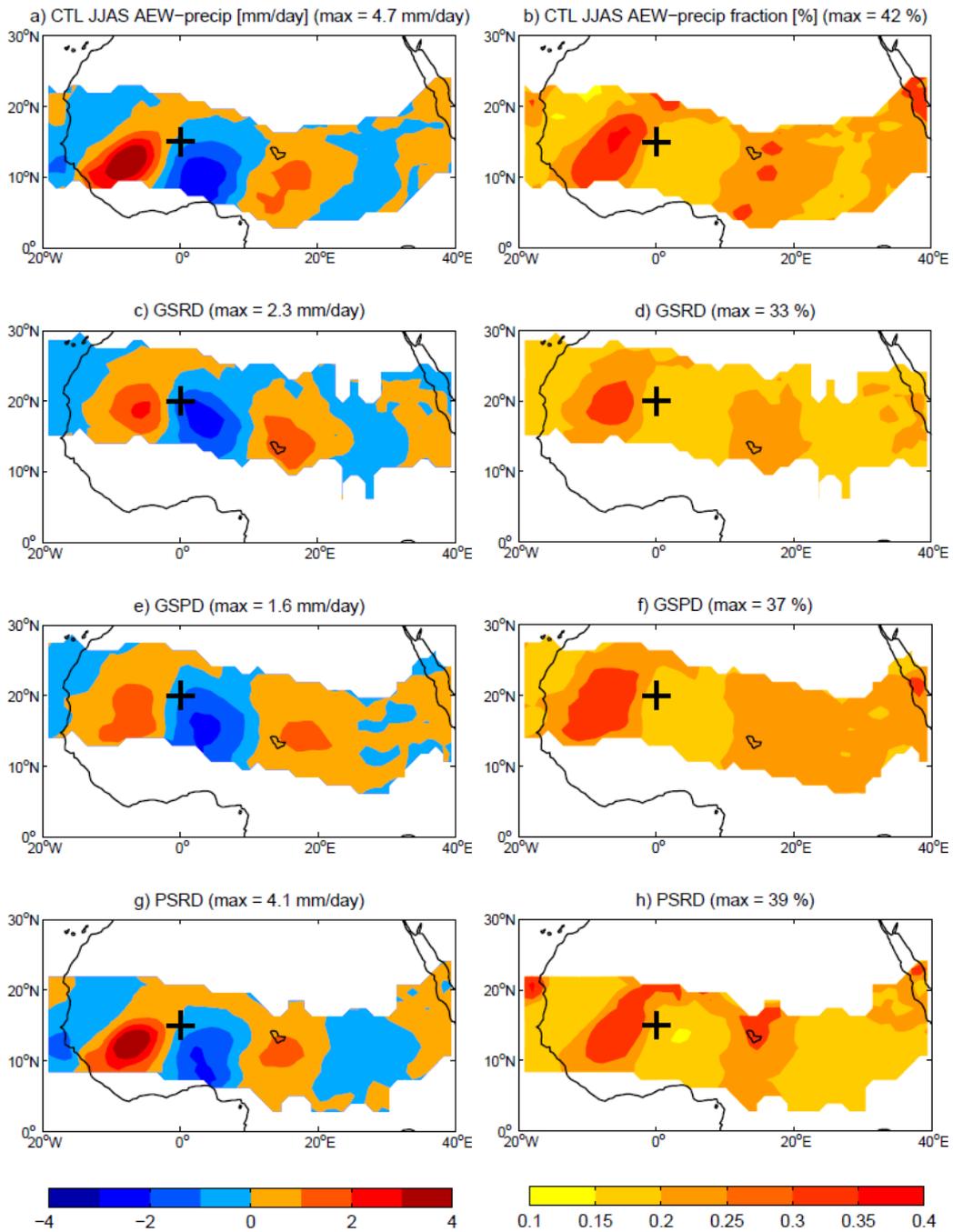
JJAS AEWs : variance of 700 hPa meridional wind (2.5–5-day filtered) [m^2/s^2]



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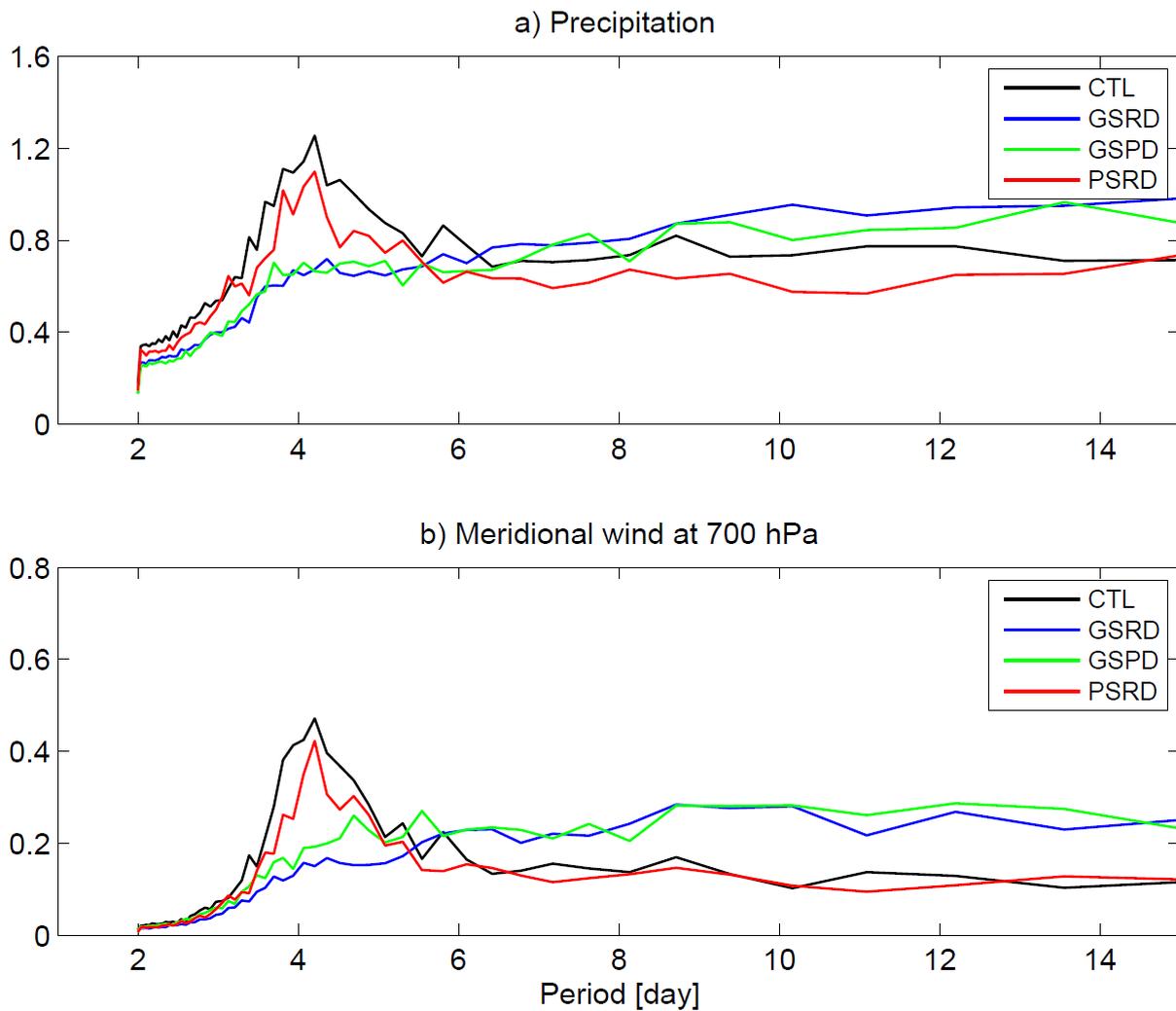
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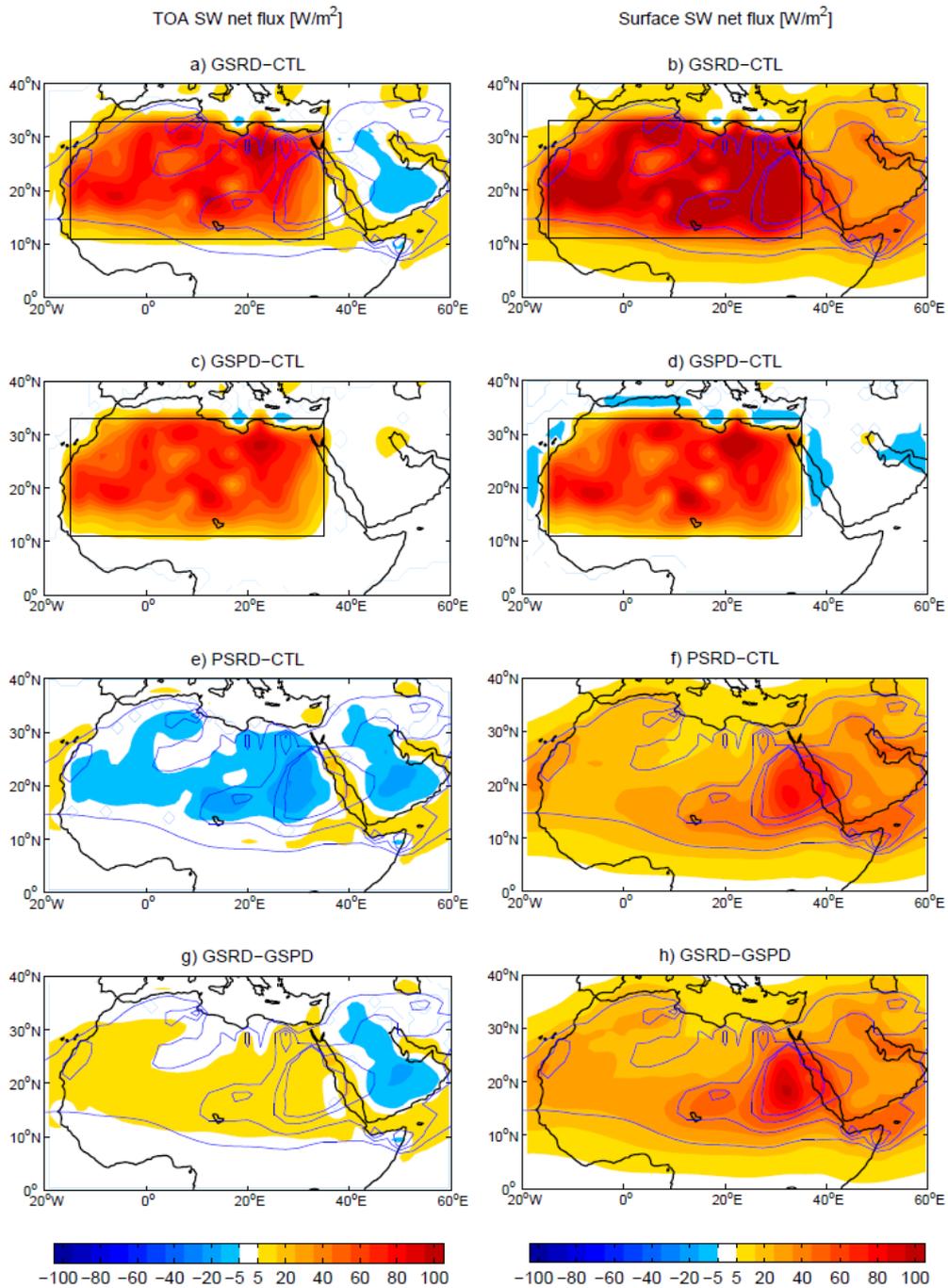
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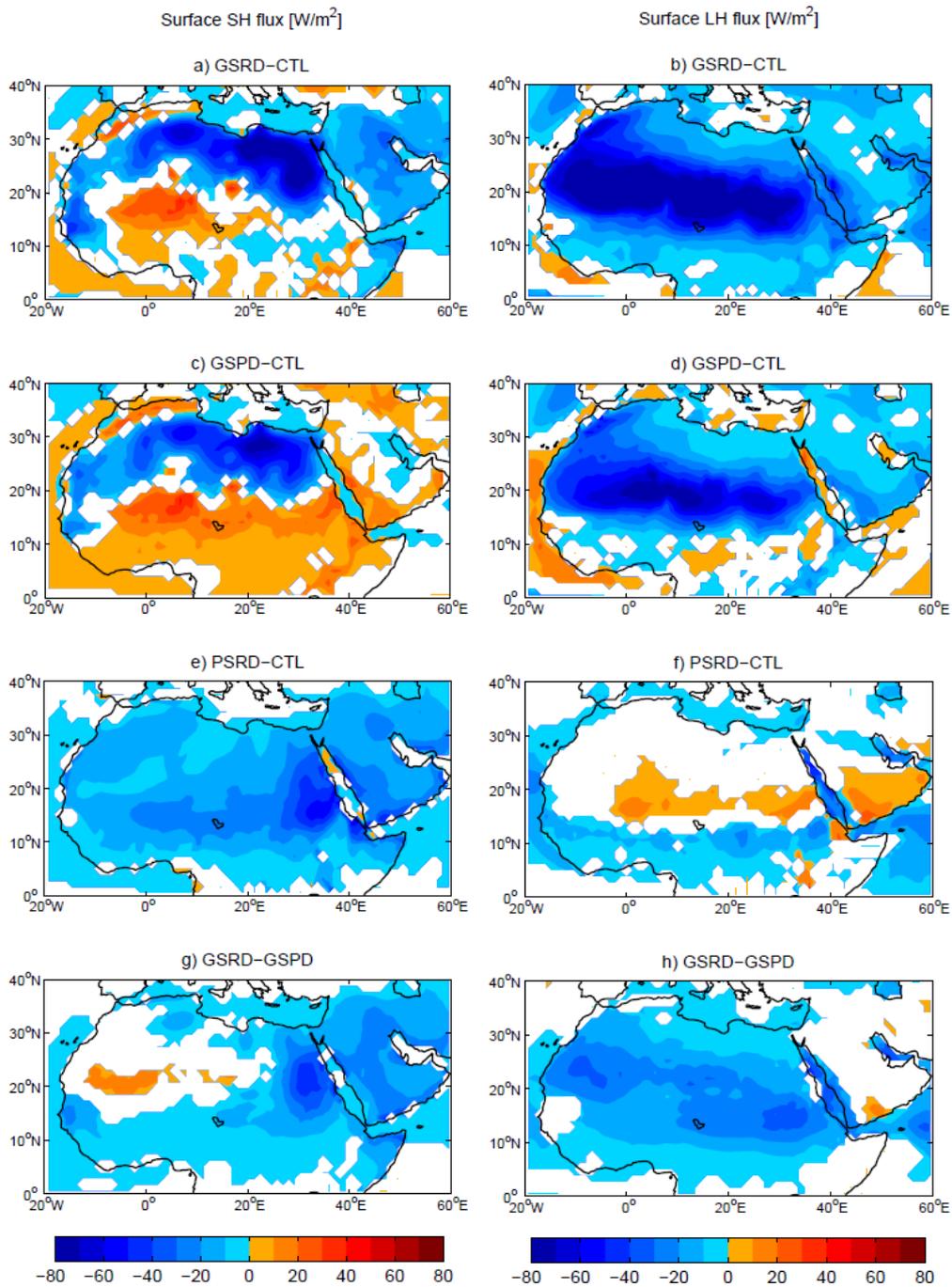


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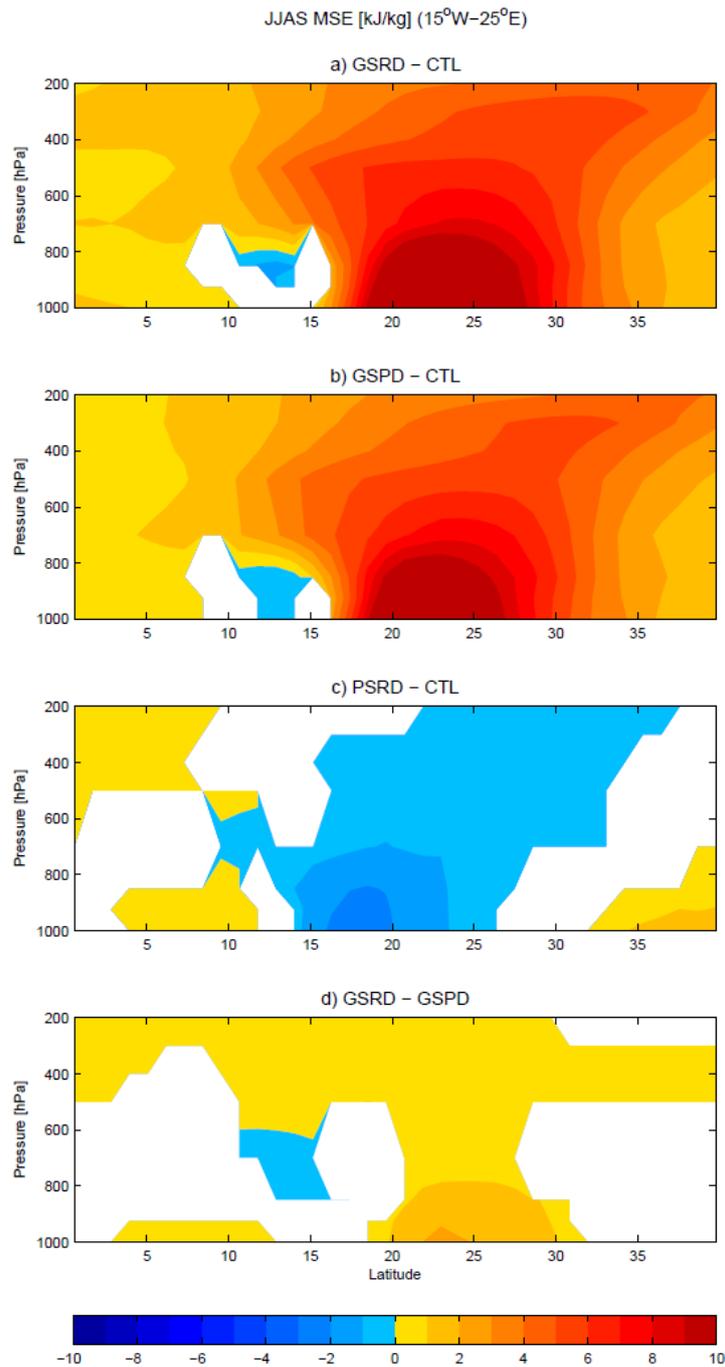
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