



**HAL**  
open science

## West African monsoon dynamics and precipitation: the competition between global SST warming and CO<sub>2</sub> increase in CMIP5 idealized simulations

Marco Gaetani, Cyrille Flamant, Sophie Bastin, Serge Janicot, Christophe Lavaysse, Frédéric Hourdin, Pascale Braconnot, Sandrine Bony

### ► To cite this version:

Marco Gaetani, Cyrille Flamant, Sophie Bastin, Serge Janicot, Christophe Lavaysse, et al.. West African monsoon dynamics and precipitation: the competition between global SST warming and CO<sub>2</sub> increase in CMIP5 idealized simulations. *Climate Dynamics*, 2017, 48 (3), pp.1353-1373. 10.1007/s00382-016-3146-z . insu-01309682

**HAL Id: insu-01309682**

**<https://insu.hal.science/insu-01309682>**

Submitted on 4 May 2016

**HAL** is a multi-disciplinary open access archive for the deposit and dissemination of scientific research documents, whether they are published or not. The documents may come from teaching and research institutions in France or abroad, or from public or private research centers.

L'archive ouverte pluridisciplinaire **HAL**, est destinée au dépôt et à la diffusion de documents scientifiques de niveau recherche, publiés ou non, émanant des établissements d'enseignement et de recherche français ou étrangers, des laboratoires publics ou privés.

1 **West African Monsoon dynamics and precipitation: the competition between global**  
2 **SST warming and CO2 increase in CMIP5 idealized simulations**

3

4 Marco Gaetani (1,2,\*), Cyrille Flamant (1), Sophie Bastin (1), Serge Janicot (3), Christophe  
5 Lavaysse (4), Frederic Hourdin (2), Pascale Braconnot (5), Sandrine Bony (2)

6

7 (1) Laboratoire Atmosphères, Milieux, Observations Spatiales, LATMOS-IPSL, UMR 8190,  
8 Sorbonne Universités (CNRS, UPMC), UVSQ, Paris, France

9 (2) Laboratoire de Météorologie Dynamique, LMD-IPSL, UMR 8539, Sorbonne Universités  
10 (CNRS, UPMC), Ecole Normale Supérieure, Ecole Polytechnique, Paris, France

11 (3) Laboratoire d'Océanographie et du Climat: Expérimentation et Approches Numériques,  
12 LOCEAN-IPSL, UMR 7159, Sorbonne Universités (CNRS, UPMC, IRD, MNHM), Paris,  
13 France

14 (4) European Commission Joint Research Centre, Institute for Environment and  
15 Sustainability, Ispra, Italy

16 (5) Laboratoire des Sciences du Climat et de l'Environnement, LSCE-IPSL, UMR 8212,  
17 CNRS, CEA, UVSQ, Gif-sur-Yvette, France

18

19 \* Correspondence: [marco.gaetani@latmos.ipsl.fr](mailto:marco.gaetani@latmos.ipsl.fr)

20

21 **Abstract**

22

23 Climate variability associated with the West African monsoon (WAM) has important  
24 environmental and socio-economic impacts in the region. However, state-of-the-art climate  
25 models still struggle in producing reliable climate predictions. An important cause of this low  
26 predictive skill is the sensitivity of climate models to different forcings. In this study, the  
27 mechanisms linking the WAM dynamics to the CO<sub>2</sub> forcing are investigated, by comparing  
28 the effect of the CO<sub>2</sub> direct radiative effect with its indirect effect mediated by the global sea  
29 surface warming. The July-to-September WAM variability is studied in climate simulations  
30 extracted from the Coupled Model Intercomparison Project Phase 5 (CMIP5) archive, driven  
31 by prescribed sea surface temperature (SST). The individual roles of global SST warming and  
32 CO<sub>2</sub> atmospheric concentration increase are investigated through idealized experiments  
33 simulating a 4 K warmer SST and a quadrupled CO<sub>2</sub> concentration, respectively. Results  
34 show opposite and competing responses in the WAM dynamics and precipitation. A dry  
35 response (-0.6 mm/day) to the SST warming is simulated in the Sahel, with dryer conditions  
36 over western Sahel (-0.8 mm/day). Conversely, the CO<sub>2</sub> increase produces wet conditions  
37 (+0.5 mm/day) in the Sahel, with the strongest response over central-eastern Sahel (+0.7  
38 mm/day). The associated responses in the atmospheric dynamics are also analysed, showing  
39 that the SST warming affects the Sahelian precipitation through modifications in the global  
40 tropical atmospheric dynamics, reducing the importance of the regional drivers, while the  
41 CO<sub>2</sub> increase reinforces the coupling between precipitation and regional dynamics. A general  
42 agreement in model responses demonstrates the robustness of the identified mechanisms  
43 linking the WAM dynamics to the CO<sub>2</sub> direct and indirect forcing, and indicates that these  
44 primary mechanisms are captured by climate models. Results also suggest that the spread in  
45 future projections may be caused by unbalanced model responses to the CO<sub>2</sub> direct and  
46 indirect forcing.

47

## 48 **1. Introduction**

49

50 West Africa is affected by large climate variability at different timescales, from interannual to  
51 multidecadal, with consequent strong environmental and socio-economic impacts, especially  
52 in the Sahelian countries, where the economy is mainly sustained by rainfed agriculture  
53 [Kandji *et al.*, 2006]. The annual precipitation in the Sahel is limited to the boreal summer  
54 season, from July to September (JAS), and it is strongly linked to the West African monsoon  
55 (WAM) dynamics [Nicholson, 2013]. After a wet period during the 50s-60s, Sahel has  
56 undergone a severe (large scale and long-lasting) drought in the 70s-80s, and a partial  
57 recovery of precipitation has been detected at the turn of the 21st century [Trenberth *et al.*,  
58 2007; Panthou *et al.*, 2014].

59

60 WAM precipitation variability at time scales from interannual to multidecadal is mainly  
61 driven by global ocean SST anomalies [Giannini *et al.*, 2003]. At the interannual time scale,  
62 the Tropical Atlantic SST variability modulates the monsoonal circulation through the  
63 meridional land-sea thermal gradient, affecting the meridional displacement of the  
64 precipitation belt [Losada *et al.*, 2010], while the Mediterranean SST variability influences  
65 the moisture transport across Sahara and the associated convergence over Sahel [Fontaine *et*  
66 *al.*, 2010; Gaetani *et al.*, 2010]. Moreover, SST variability in the tropical Pacific may induce  
67 stationary Kelvin and Rossby waves propagating along the Equator and interacting over the  
68 Sahel [Rowell, 2001]. The interannual regional and remote SST connections are not  
69 stationary and are modulated by decadal and multidecadal SST anomalies [Fontaine *et al.*,  
70 2011]. The global ocean signals which dominate the low frequency precipitation variability in  
71 the Sahel are: the tropical SST warming, the Atlantic Multidecadal Variability (AMV), and  
72 the Interdecadal Pacific Oscillation (IPO). The tropical warming, associated to global  
73 warming and IPO positive phases, favours dry conditions in the Sahel, through the inhibition  
74 of the tropical convection [Bader and Latif, 2003; Lu and Delworth, 2005; Villamayor and  
75 Mohino, 2015]. On the other hand, positive phases of AMV, by displacing northward the  
76 intertropical convergence zone (ITCZ), favour precipitation in the Sahel [Zhang and  
77 Delworth, 2006; Ting *et al.*, 2009; Mohino *et al.*, 2011]. A broad and up-to-date review of the  
78 literature on the SST influence on the climate variability in West Africa is reported by  
79 [Rodríguez-Fonseca *et al.*, 2015].

80

81 The WAM variability is modulated also by atmospheric circulation patterns at the regional  
82 scale, and a key driver is the thermal depression associated with the surface heating over  
83 West Africa in boreal summer. *Lavaysse et al.* [2009] described the seasonal evolution of this  
84 thermal low, identifying the Saharan heat low (SHL) as the area of maximum expansion of  
85 the lower troposphere, which in summer is located over Sahara, west of 10°E. The SHL onset  
86 is closely linked to the WAM onset in late June [*Lavaysse et al.*, 2009], and the monsoonal  
87 precipitation evolution from the onset to its mature stage is closely related to the SHL  
88 behaviour [*Xue et al.*, 2010], with strong SHL phases associated to atmospheric circulation  
89 patterns favourable to wet conditions over central Sahel, and dry anomalies over Senegal  
90 [*Lavaysse et al.*, 2010b]. The SHL is sensitive to regional forcings, such as synoptic systems  
91 travelling across the Sahel [*Lavaysse et al.*, 2010a], dust emission from Sahara and Sahel  
92 [*Lavaysse et al.*, 2011], and local water-vapour-temperature feedback [*Evan et al.*, 2015]. On  
93 the other hand, the SHL variability is modulated by mid-latitude circulation synoptic  
94 disturbances [*Chauvin et al.*, 2010], and northerly advections of cold air and moisture from  
95 the Mediterranean region [*Vizy and Cook*, 2009; *Gaetani et al.*, 2010].

96

97 In the last 15 years, a big effort has been made in the framework of the Coupled Model  
98 Intercomparison Project Phase 3 and 5 (CMIP3 [*Meehl et al.*, 2007] and CMIP5 [*Taylor et*  
99 *al.*, 2012], respectively) to model climate variability in West Africa, with promising but still  
100 unsatisfying results. Specifically, state-of-the-art coupled climate models show poor ability in  
101 correctly capturing the observed WAM precipitation variability, in terms of amplitude,  
102 phases and trends [*Biasutti*, 2013], mainly because of models' low skill in reproducing the  
103 observed SST teleconnections [*Rowell*, 2013]. Furthermore, in both CMIP3 and CMIP5  
104 simulations, sizable uncertainties affect future projections, with a large spread among the  
105 outputs [*Biasutti*, 2013]. Most of the models predict wet conditions in the 21st century in the  
106 Sahel, which are explained through the direct effect of the increase in green-house gases  
107 (GHG) concentration on net radiation at the surface, which in turn leads to local increased  
108 evaporation and vertical instability [*Haarsma et al.*, 2005; *Hoerling et al.*, 2006; *Giannini*,  
109 2010]. On the other hand, some models project dry conditions as a consequence of the global  
110 ocean warming, which heats the troposphere and imposes stability, reducing moisture  
111 transport and deep convection over land [*Held et al.*, 2005; *Cook and Vizy*, 2006; *Caminade*  
112 *and Terray*, 2010]. In this respect, *Bony et al.* [2013] pointed out that about half the tropical  
113 circulation change projected by the end of the 21st century, and consequently a large fraction  
114 of the regional precipitation change, is independent of global surface warming, but it is rather

115 a response to the weaker net radiative cooling of the atmosphere associated with higher  
116 atmospheric CO<sub>2</sub> levels affecting the strength of the atmospheric vertical motions. Therefore,  
117 the competition between the response of the land-atmosphere system to the local GHG  
118 radiative forcing, and the response mediated through the warming of the global SST  
119 [*Giannini, 2010; Skinner et al., 2012*], emerges as a fundamental component of the WAM  
120 dynamics and variability, and different sensitivities of climate models to each process appear  
121 as a possible explanation of the spread in the projections. Moreover, *Biasutti et al. [2009]*  
122 highlighted that the relationship between the regional atmospheric circulation, particularly the  
123 thermal low over the Sahara desert, and Sahel precipitation is generally well reproduced by  
124 the coupled models, both in the 20th and the 21st century, while the same models show  
125 discrepancies in future projections. The authors argue that the spread in predictions could be  
126 also explained through different abilities in reproducing the mechanisms which influence the  
127 regional atmospheric circulation.

128

129 In this context, the open issues regarding the WAM dynamics and precipitation variability  
130 can be conveyed in two main arguments. The first one concerns the detailed comprehension  
131 of the physical mechanisms connecting the direct and indirect GHG forcing to the WAM  
132 dynamics, and the specific roles of the large scale and regional atmospheric dynamics in  
133 modulating this linkage. The second one concerns the assessment of the ability of climate  
134 models to simulate the identified physical mechanisms. The objective of this study is to  
135 address these issues, by investigating the specific influence of the CO<sub>2</sub> atmospheric  
136 concentration on the WAM dynamics. The direct effect of CO<sub>2</sub> concentration increase, and  
137 its indirect effect mediated by the global SST warming, are analysed, with a specific focus on  
138 the atmospheric dynamics response at the regional and global scale. To this aim, multi-model  
139 SST-driven sensitivity experiments are analysed, simulating the present climate under  
140 idealized warmer Global Ocean and higher CO<sub>2</sub> atmospheric concentration, respectively. To  
141 allow a clear separation of the climate system response, and assess the model sensitivity to  
142 the diverse forcings, extreme conditions are imposed in the sensitivity experiments. SST-  
143 driven simulations are preferred to coupled simulations to minimize the problems affecting  
144 the coupled models in correctly locating the Sahelian monsoonal precipitation, due to SST  
145 biases in the equatorial Atlantic [*Roehrig et al., 2013*].

146

147 The paper is structured as follow: models and data analysed are presented in Section 2; the  
148 response of monsoonal precipitation and circulation in the idealized experiments is presented

149 in Section 3; the modifications in the global and regional dynamics associated to the WAM  
150 are discussed in Section 4 and 5, respectively; finally, conclusions are drawn in Section 6.

151

## 152 **2. Model simulations**

153

154 The JAS WAM variability and dynamics are studied by analysing numerical experiments  
155 from a set of 12 state-of-the-art atmospheric-ocean coupled models selected in the CMIP5  
156 archive [Taylor *et al.*, 2012] (see Table 1 for details on the models). In the control simulation  
157 (CTL), the models are run in atmospheric-only configuration, with prescribed observed SST  
158 and sea ice for the period 1979 to the present. Moreover, the simulation includes the observed  
159 evolution in the atmospheric composition (including CO<sub>2</sub>), due to both anthropogenic and  
160 natural influences; and the changes in solar forcing, emissions or concentrations of aerosols,  
161 and land use. The CTL simulation is the baseline for two sensitivity experiments run either by  
162 prescribing uniform 4 K increase in global SST (4K experiment), or by quadrupling the CO<sub>2</sub>  
163 atmospheric concentration while maintaining the SST unchanged. (4xCO<sub>2</sub> experiment). The  
164 experimental setup is described in detail in Taylor *et al.* [2012]. Imposing extreme idealized  
165 forcing in the sensitivity experiments allows to maximise the impacts and extract the slow  
166 response of the climate system to the global SST warming (in 4K), and the fast response to  
167 the local direct CO<sub>2</sub> radiative forcing (in 4xCO<sub>2</sub>), respectively. The 4xCO<sub>2</sub> and 4K idealized  
168 forcings can be considered extreme in comparison with the observed changes in CO<sub>2</sub>  
169 concentration (+40% since 1750) and SST (+0.7 K in the period 1880-2012) from the pre-  
170 industrial through the historical period [IPCC, 2014]. However, the idealized scenarios used  
171 in this study are comparable with the situation expected in 2100 in the RCP8.5 emission  
172 scenario, with the CO<sub>2</sub> concentration augmented from 390 ppm in 2011 to more than 1000  
173 ppm (more than +260%), and more than 3 K global SST warming [IPCC, 2014]. RCP8.5 is  
174 an extreme scenario with a radiative forcing reaching 8.5 W/m<sup>-2</sup> in 2100 [Riahi *et al.*, 2011].

175

176 To ensure that multi-model mean results are not biased towards models for which more  
177 realisations are available, for each model only one realisation of each experiment is used.  
178 This “one model-one vote” concept [Santer *et al.*, 2009] has been proved successful for  
179 detection of Sahel rainfall significant anomalies in CMIP3 and CMIP5 climate scenarios  
180 [Monerie *et al.*, 2012]. The selection of models is based on the experiments availability,  
181 considering only models for which the three experiments are available. The differences in  
182 grid resolution among models are harmonised by regridding all the datasets to a T42 (~2.8°)

183 Gaussian grid. The simulations are analysed over a 30-year period from 1979 to 2008, and  
184 averaging precipitation and atmospheric fields in JAS. The response of the WAM  
185 precipitation and dynamics to idealized conditions is studied by comparing the sensitivity  
186 experiments to the CTL simulation, and significant differences in means are detected through  
187 a Student's t-test at 95% confidence level. A comprehensive evaluation of the individual  
188 models' performances is beyond the scope of the paper. However, the biases of the models  
189 for some variables describing the WAM dynamics are presented and discussed in the  
190 Supplementary Material, along with the response of the individual models in the sensitivity  
191 experiments.

192

### 193 **3. Monsoonal circulation and precipitation response**

194

195 In Figure 1, the observed and MMM precipitation fields are presented. Precipitation  
196 simulated in the CTL experiment is compared to the Global Precipitation Climatology Project  
197 version 2.2 (GPCP) dataset [Adler *et al.*, 2003], which contains a monthly analysis of global  
198 surface precipitation at 2.5° resolution from January 1979 to the present. GPCP analysis  
199 merges precipitation estimates from satellite products and surface rain gauge observations,  
200 and it shows good reliability at seasonal time scale and better performances than satellite-  
201 only products [Nicholson *et al.*, 2003]. Moreover, GPCP has the clear advantage of extending  
202 the precipitation analysis to the ocean, allowing a better comparison with model simulations.  
203 The MMM of the CTL simulation (Figure 1b) reproduces the main spatial structures of the  
204 observed WAM precipitation (Figure 1a), with a general underestimation of the seasonal  
205 mean (Figure 1c). The precipitation amount averaged in the domain [5°-20°N, 20°W-30°E]  
206 including the JAS precipitation belt over West Africa, is 5.1 mm/day in the GPCP dataset,  
207 with 0.5 mm/day standard deviation (STD). In the CTL simulation, the MMM is 4.1 mm/day,  
208 around 1 mm/day lower than the observed precipitation, and the multi-model spread ranges in  
209 [2.4, 6.7] mm/day, with 1.2 mm/day STD. Moreover, the CTL simulation is strongly biased  
210 in proximity of the climatological rainfall maxima, over western Sahel (-3.8 mm/day around  
211 10°N, 15°W) and Nigeria and Cameroun (-6.6 mm/day around 5°N, 10°E), revealing a  
212 general inability of the models to fully capture the spatial variability of precipitation (see also  
213 individual model biases presented in the Supplementary Material, Figure S1). The origin of  
214 this limitations may be attributed to shortcomings in the model physics, such as the  
215 inaccurate representation of clouds, energy fluxes, coupling at the surface, and convection at  
216 the sub-grid scale, but also to the coarse resolution of the global climate models, which is



217 insufficient to simulate the mesoscale convection characterizing the WAM precipitation.  
218 These biases underline the urgent need for improvement in the simulation of the  
219 climatological WAM precipitation. However, they do not compromise the objective of this  
220 study, which is focused on the understanding of the WAM dynamics response in idealized  
221 climate conditions, rather than on making quantitative predictions.  
222

223 Results for 4K experiment (Figure 1d) show a dry response in the Sahel to SST warming,  
224 with dryer conditions over western Sahel. On the contrary, 4xCO<sub>2</sub> experiment (Figure 1e)  
225 shows wet conditions in the Sahel, with the strongest response to CO<sub>2</sub> increase over central-  
226 eastern Sahel. These responses to idealized conditions are in line with previous results on the  
227 influence of SST warming [*Held et al.*, 2005; *Cook and Vizy*, 2006; *Caminade and Terray*,  
228 2010] and GHG increase [*Haarsma et al.*, 2005; *Hoerling et al.*, 2006; *Giannini*, 2010] on  
229 WAM precipitation. The modifications in the precipitation patterns are detailed for the  
230 individual models by computing a regional rainfall index for the Sahel, the Sahelian rainfall  
231 index (SRI) in the domain [10°-20°N, 20°W-30°E], and two sub-regional indices for western  
232 [20°W-0°E] and eastern [0°-30°E] Sahel. In the 4K simulation (Figure 2a), all the models  
233 show significant negative SRI differences with CTL, the MMM difference is -0.6 mm/day,  
234 and the multi-model spread ranges between -1.1 and -0.3 mm/day (0.3 mm/day STD). In  
235 western Sahel (Figure 2b), the 4K MMM response decreases to -0.8 mm/day, and differences  
236 are significant in all the models, comprised in [-1.7, -0.2] mm/day (0.5 mm/day STD). In  
237 eastern Sahel (Figure 2c), differences are significant in 10 out of 12 models, ranging in [-1.0,  
238 0.0] mm/day (0.3 mm/day STD), with a -0.5 mm/day MMM difference. Conversely, the SRI  
239 response in the 4xCO<sub>2</sub> MMM is +0.5 mm/day, the multi-model spread ranges from +0.2 to  
240 +0.9 mm/day (0.2 mm/day STD), and the differences are significant in 11 out of 12 models  
241 (Figure 2a). In western Sahel (Figure 2b), the 4xCO<sub>2</sub> MMM response decreases to +0.3  
242 mm/day, and differences are significant in 5 out of 12 models, ranging in [0.0, +0.8] mm/day  
243 (0.3 mm/day STD). In eastern Sahel (Figure 2c), the 4xCO<sub>2</sub> MMM response increases to  
244 +0.7 mm/day, and differences are significant in 11 out of 12 models, ranging in [+0.2, +1.2]  
245 mm/day (0.3 mm/day STD). The substantial multi-model agreement on the sign of the  
246 precipitation responses confirms the robustness of the simulated precipitation patterns in  
247 response to the SST and CO<sub>2</sub> idealized forcings. However, the spread in the multi-model  
248 response suggests possible higher sensitivity in individual models to one of the two forcings.  
249 The relative importance of the SST warming and CO<sub>2</sub> increase in influencing the Sahelian  
250 precipitation is assessed by computing the sensitivity of the SRI to the global surface

251 warming induced by the idealized forcings, i.e., the ratio between the changes in SRI and  
252 global surface temperature averaged between 50°S and 70°N. It results that, in the 4xCO<sub>2</sub>  
253 experiment, sensitivity is +1.27 mm/day/K in the MMM. This is, in absolute value, 9 times  
254 higher than in 4K experiment, showing -0.14 mm/day/K in the MMM. A prominent influence  
255 of the direct CO<sub>2</sub> forcing on the Sahelian precipitation is evident, although differences in the  
256 individual model sensitivity are sizable, ranging from +0.53 mm/day/K in FGOALS-g2 (3  
257 times higher than in the 4K experiment, -0.16 mm/day/K) to +2.17 mm/day/K in HadGEM2-  
258 A (24 times higher than in the 4K experiment, -0.09 mm/day/K).

259

260 In addition, it is interesting to notice how, from the superimposition of the responses to SST  
261 warming and CO<sub>2</sub> increase, the main features of the projected changes in Sahelian  
262 precipitation emerge, namely dry anomalies over western Sahel and wet conditions over  
263 central-eastern Sahel [*Biasutti, 2013; Monerie et al., 2013; Roehrig et al., 2013*]. A  
264 descriptive illustration is presented in Figure 3, where the 4K-CTL and 4xCO<sub>2</sub>-CTL  
265 precipitation differences (Figures 1d, e) are combined and compared with the 21st century  
266 projection simulated by the same models, run in the coupled setup, for the RCP8.5 emission  
267 scenario. The combination of the 4K and 4xCO<sub>2</sub> responses (Figure 3a, computed as the sum  
268 of the precipitation patterns displayed in Figures 1d, e) exhibits positive precipitation  
269 anomalies in central Sahel (up to around +0.5 mm/day) and the Tropical Atlantic around 5°N  
270 (up to around +1 mm/day), and dry anomalies in western Sahel (down to around -2 mm/day)  
271 and the eastern part of the Guinean coast (down to around -0.5 mm/day). In the RCP8.5  
272 scenario, at the end of the 21st century, wet anomalies are expected in central-eastern Sahel  
273 (up to around +1 mm/day) and the Tropical Atlantic around 5°N (up to around +2 mm/day),  
274 and dry anomalies are projected in western Sahel (down to around -0.5 mm/day) and the  
275 Guinea coast (down to around -1.5 mm/day). Whereas the anomaly patterns are different in  
276 amplitude and local features, the main spatial structures appear similar. This suggests that the  
277 future projected pattern may result from the competition between the different actions of the  
278 CO<sub>2</sub> forcing on the land-atmosphere system. On the one hand, the indirect action conveyed  
279 by the SST warming, which heats the troposphere, imposing stability and reducing deep  
280 convection over land [*Caminade and Terray, 2010*]. On the other hand, the direct action on  
281 the local radiative balance, producing a weaker net radiative cooling of the atmosphere, a  
282 warming of the land surface and a change in the land-sea thermal gradient, which is  
283 favourable to the strengthening of large-scale upward motions [*Bony et al., 2013*]. The  
284 evidence of the contrasting effects of the CO<sub>2</sub> forcing on the WAM precipitation, along with

285 the different sensitivity shown by individual models, point out the importance of  
286 understanding the balance between the competing actions of the SST warming and CO<sub>2</sub>  
287 increase and improving the climate model ability in correctly simulate this balance. It is also  
288 worth noticing that the superimposition pattern in Figure 3 is consistent with the precipitation  
289 anomalies expected during positive phases of the SHL [Lavaysse *et al.*, 2010b; Evan *et al.*,  
290 2015].

291  
292 The monsoonal flow in the lower troposphere is depicted in Figure 4 through the horizontal  
293 wind field and the divergence of the moisture transport at 925 hPa. The CTL simulation  
294 shows the intertropical discontinuity (ITD) between the southwesterly moist flow from the  
295 tropical Atlantic and the northerly dryer flow crossing the Sahara [Issa Lélé and Lamb, 2010],  
296 and the associated moisture convergence over the Sahel (Figure 4a). The response of the  
297 models to the SST warming is a general weakening of the monsoonal circulation, with no  
298 significant changes over the Sahel, accompanied by a slight southward displacement of the  
299 moisture transport convergence (Figure 4b). The shift of the latitudinal position of the  
300 moisture transport convergence over West Africa is assessed by identifying the mean position  
301 of the divergence minimum in the longitudinal domain [10°W-30°E]. The shift in the  
302 minimum is -0.5° in the MMM, it is negative in 9 out of 12 models, and the multi-model  
303 spread ranges between -1.8° and +1.0° (0.7° STD). Given the small modifications in the wind  
304 field, the changes in the moisture convergence can be attributed to the general moistening of  
305 the lower troposphere produced by the SST warming (not shown). On the contrary, the  
306 massive injection of CO<sub>2</sub> in the atmosphere produces a sizable strengthening of the westerly  
307 flow from the Atlantic across the Sahel, and a slight reinforcement of the northerly flow from  
308 the Mediterranean, resulting in a clear northward displacement of the moisture convergence  
309 belt over the Sahel (Figure 4c). The shift is +1.1° in the MMM, is positive in 11 out of 12  
310 models, ranging in [0.0°, +3.6°] (0.9° STD).

311

#### 312 **4. Response at the global scale**

313

314 The competition between the direct and indirect effects of CO<sub>2</sub> concentration increase in  
315 modifying circulation and precipitation in the global Tropics has been investigated by Bony *et al.*  
316 [2013]. By separating the precipitation changes in a dynamic component related to  
317 modifications in large scale circulation and a thermodynamic one independent of these, they  
318 demonstrated that about half the changes projected in tropical circulation and regional

319 precipitation by the end of the 21st century is independent of global surface warming. In  
320 Figure 5, the global net radiative balance, defined at the top of the atmosphere as the  
321 difference between the downward shortwave radiation and the sum of upward shortwave and  
322 longwave radiation, is presented for CTL and sensitivity experiments. The CTL experiment  
323 (Figure 5a) represents the warming of the atmosphere in the Tropics and sub-Tropics, with  
324 cooling over the desert belt in the Northern Hemisphere, and in the extra-Tropics (north of  
325 45-50°N and south of 10-15°S). In the 4K simulation (Figure 5b), a generalized net radiative  
326 cooling of the atmosphere is found, reaching  $-16 \text{ W/m}^2$  in the Sahara, which is ascribed to the  
327 increase in losses of longwave radiation associated to the surface warming (not shown). On  
328 the contrary, the increased CO<sub>2</sub> concentration produces a net radiative warming in 4xCO<sub>2</sub>  
329 (Figure 5c), reaching  $+14 \text{ W/m}^2$  in the Sahara, which is associated with reduced losses in  
330 both longwave and shortwave radiation (not shown).

331

332 The response in the lower troposphere to the changes in the radiative balance is studied  
333 through the moist and dry static energy (MSE and DSE, respectively) content in the lower  
334 troposphere integrated between 925 and 700 hPa. The MSE is defined as

$$335 \quad MSE = gz + C_p T + Lq,$$

336 where  $gz$  is the geopotential energy (with  $g$  the gravitational acceleration and  $z$  the  
337 geopotential height),  $C_p T$  the enthalpy (with  $C_p$  the specific heat of dry air at constant  
338 pressure and  $T$  the temperature), and  $Lq$  is the latent energy associated with evaporation and  
339 condensation of water (with  $L$  the latent heat of evaporation and  $q$  the specific humidity). As  
340 transformation of enthalpy and latent energy available in the lower troposphere into  
341 geopotential energy in the upper levels is the main signal of convection, MSE is directly  
342 related to monsoonal precipitation [*Fontaine and Philippon, 2000*]. The DSE accounts for  
343 only enthalpy and geopotential energy. The differences between sensitivity and CTL  
344 experiments are presented in Figure 6. The CTL simulation shows a maximum in the tropical  
345 belt, which is associated to the latent heat content, while DSE is more important over the  
346 subtropical continental surfaces (Figure 6a, b). The surface warming in 4K produces a global  
347 and almost uniform increase of MSE, with a maximum over the northern Tropics, which is  
348 associated to the changes in the latent heat content (Figure 6c, d). On the contrary, the  
349 response to increased CO<sub>2</sub> concentration in 4xCO<sub>2</sub>, although weaker, shows a clear  
350 relationship with the changes in the DSE content, resulting in a meridional gradient and well  
351 localized maxima over the Sahel, the Arabian Peninsula, southwestern and eastern Asia, and  
352 North America (Figure 6e, f). It appears that the response of MSE to the SST warming is

353 mostly associated to the change of moist component at the global scale, while the response in  
354 the 4xCO<sub>2</sub> experiment is driven by the modifications in the dry component localized over the  
355 northern continental masses.

356

357 The main features of the 3D large scale circulation affecting the WAM variability are studied  
358 by analysing the meridional circulation, displayed through meridional wind and the vertical  
359 velocity averaged in the domain [10°W, 10°E] to include southern Atlantic Ocean and West  
360 Africa (Figure 7), and the zonal circulation, displayed through zonal wind and vertical  
361 velocity averaged in the domain [5°N, 20°N], to include Sahel and northern Tropics (Figure  
362 8).

363

364 The CTL simulation captures the main features of the meridional circulation over southern  
365 Atlantic and West Africa, illustrating the northern and southern overturning cells associated  
366 to the meridional Hadley circulation, along with the deep convection associated to the ITCZ  
367 around 10°N and the convergence in the lower troposphere over the Sahel associated to the  
368 monsoonal circulation (Figure 7a). By comparing CTL with the sensitivity experiments, it is  
369 seen that the SST warming weakens the southern overturning cell, resulting in a reduction of  
370 the deep convection over Guinea and Sahel (Figure 7b). On the contrary, in 4xCO<sub>2</sub> the  
371 southern overturning cell is reinforced along with deep convection over the Sahel, and the  
372 lower troposphere convergence move northward into the Sahara (Figure 7c).

373

374 The zonal circulation over the northern Tropics is dominated by upward motions associated  
375 with the WAM dynamics and induced by the oceanic warm pools in the Indian Ocean and in  
376 western and eastern Pacific (Figure 8a). Moreover, the Tropical easterly jet (TEJ) axis is  
377 recognizable in the upper troposphere over Africa and the Indian Ocean, along with the  
378 westerly flow in the lower troposphere over the Indian Ocean associated to the Indian  
379 summer monsoon (Figure 8a). The TEJ intensity is strongly related to the precipitation  
380 variability, wet conditions in the Sahel being concomitant with stronger than average TEJ and  
381 vice versa [Nicholson, 2013]. The sensitivity experiments show anomalous ascending  
382 motions as a response to SST warming over the western Pacific and anomalous subsidence  
383 over Africa and Indian Ocean, along with a weakening of the TEJ (Figure 8b). The change in  
384 the TEJ intensity is assessed over West Africa by averaging the zonal wind at 200 hPa in the  
385 domain [5°-15°N, 20°W-30°E]. It results that, in the 4K MMM, the TEJ is 4.7 m/s weaker  
386 than in the CTL run, and the multi-model response ranges in [+0.5, +10.1] m/s (2.8 m/s

387 STD). Conversely, the 4xCO<sub>2</sub> experiment simulates weaker upward motions in the western  
388 and eastern Pacific, and an anomalous organized zonal cell over the Africa-Atlantic sector,  
389 with a westerly flux in the lower troposphere into West Africa (see also Figure 4c) and a  
390 stronger TEJ (Figure 8c). In the 4xCO<sub>2</sub> experiment, the MMM response is -2.9 m/s, and the  
391 multi-model response ranges in [-3.75, -0.12] m/s (0.97 m/s STD).

392

393 The response of the teleconnection patterns is explored through the global velocity potential  
394 and associated divergent wind at 200 hPa (Figure 9). The CTL simulation shows a  
395 divergence-convergence dipole between western Pacific and southern Atlantic (Figure 9a),  
396 which is the dominant teleconnection pattern in boreal summer [Tanaka *et al.*, 2004]. The  
397 sensitivity experiments confirm the responses seen in the zonal circulation, with a general  
398 weakening of the global teleconnection pattern in 4K (Figure 9b), and a divergence-  
399 convergence anomaly between tropical Africa and Atlantic, indicating an intensification of  
400 the teleconnection over Africa in 4xCO<sub>2</sub> (Figure 9c).

401

402 The analysis of the response of the global dynamics fields and the modifications in the  
403 Sahelian precipitation connections suggests that the increase in CO<sub>2</sub> concentration favours  
404 the meridional circulation and the zonal connectivity in the tropical belt, by forcing a  
405 meridional energy gradient. Meridional circulation and zonal connectivity are conversely  
406 weakened by a warmer Global Ocean, which produces an energy excess in the tropical belt.

407

## 408 **5. Response of the regional drivers**

409

410 In this section the impact of the SST and CO<sub>2</sub> idealized forcings on the WAM regional  
411 drivers is discussed, by analysing the modifications in the most relevant features of the  
412 atmospheric circulation over West Africa, and the changes in their relationship with the  
413 Sahelian precipitation.

414

415 The response of the SHL is first analysed, by computing the 925-700 hPa thickness ( $Z_{925-700}$ ), which measures the thermal expansion of the lower troposphere [Lavaysse *et al.*, 2009]  
416 The CTL simulation well reproduces the climatological location of SHL over western Sahara  
417 (Figure 10a). The 4K experiment is characterized by a general thickening of the lower  
418 troposphere compared to CTL (Figure 10b), which is consistent with the global warming of  
419 the surface associated with the forcing SST pattern (not shown). A peak in thickness is  
420

421 evident along 15°N (around +60 m), which inhibits the penetration of the monsoonal flow  
422 and rainfall inland compared to CTL (see Figure 1d). The SHL latitudinal position is  
423 identified as the mean position of the Z925-700 maximum in the longitudinal domain [10°W,  
424 10°E]. In the 4K simulation, 10 out of 12 models show a southward shift, and the multi-  
425 model spread ranges in [-2.0°, +0.6°], (0.7° STD), resulting in a -0.7° meridional  
426 displacement in the MMM. On the contrary, the 4xCO<sub>2</sub> experiment presents a significant  
427 warming over land in the boreal summer hemisphere, resulting in thermal anomalies over the  
428 continental surface in North America, Eurasia and the Sahara (not shown). As a consequence  
429 of the Sahara desert warming, thickness over West Africa shows a meridional gradient (from  
430 around 0 m in the Sahel to around +17 m along the Mediterranean southern coast, see Figure  
431 10c), which is favourable to the northward displacement of the precipitation belt (see Figure  
432 1e). The SHL displacement is +1.0° northward in the 4xCO<sub>2</sub> MMM, comprised in [0.0°,  
433 +2.6°] (0.9° STD) in the multi-model ensemble. The comparison between 4K and 4xCO<sub>2</sub>  
434 changes in thickness shows a larger thickening in 4K everywhere in West Africa, which is  
435 associated with reduced precipitation, pointing to the importance of the regional pattern of the  
436 modifications, in spite of the mere amplitude.

437

438 The modifications in the MSE and DSE content in the lower troposphere are detailed for  
439 West Africa by computing the meridional profile in the domain [10°W-10°E], to emphasize  
440 the land-sea contrast between West Africa and southern Atlantic (Figure 11). Differences  
441 between sensitivity experiments and CTL show a general and significant MSE increase in all  
442 the 4K models, from tropical Atlantic to Sahara and peaking over the coast of Guinea (Figure  
443 11a). On the contrary, the modifications simulated in the 4xCO<sub>2</sub> experiment are in general  
444 smaller, with all the models simulating significant and positive differences over the African  
445 continent and peaking over the Sahel, and no significant differences over the ocean (Figure  
446 11b). When DSE is selected (Figure 11c, d), it results that the meridional profile is dominated  
447 in both the experiments by the latent heat associated with evaporation. Enthalpy over Sahara  
448 desert is important in maintaining the MSE meridional gradient in 4xCO<sub>2</sub>, but in 4K it is not  
449 enough to produce a significant meridional gradient. In spite of the larger availability of MSE  
450 in the 4K experiment compared to 4xCO<sub>2</sub>, the simulated profile is less favourable to the  
451 inland progression of the monsoonal circulation, because of the peak around the Guinea  
452 coast, which prevents the monsoonal flow to penetrate inland (Figure 11c). In comparison  
453 with the CTL simulations, the maximum of the MSE meridional profile is shifted 1.2°  
454 southward in the MMM, comprised between -2.4° and 0.0° (0.7° STD) in the multi-model

455 ensemble. On the contrary, the 4xCO<sub>2</sub> profiles show a clear meridional gradient from the  
456 Atlantic to the Sahel, which is more favourable to the monsoonal dynamics (Figure 11d).  
457 The MSE maximum results northward shifted by 0.7°, ranging in [+0.0°, +1.6°] (0.5° STD) in  
458 the multi-model ensemble. Furthermore, this enhanced energy meridional gradient is also  
459 consistent with the anomalous zonal circulation over West Africa displayed in Figure 7c.

460

461 The zonal wind at 600 hPa is analysed to portray the African easterly jet (AEJ), which is a  
462 key element in the WAM dynamics. The AEJ develops in response to the meridional thermal  
463 gradient over West Africa, with an east-west orientation along 12°N [Cook, 1999], and is  
464 maintained by the meridional circulations associated with the monsoon, namely the Hadley  
465 circulation and the SHL [Thorncroft and Blackburn, 1999; Chen, 2005]. The AEJ variability  
466 influences the intensity and position of the WAM precipitation pattern, with a stronger  
467 (weaker) than average jet associated to negative (positive) rainfall anomalies over Sahel, and  
468 a northern (southern) displacement associated to a more (less) inland penetrating monsoonal  
469 flow [Cook, 1999]. Moreover, AEJ may modulate the WAM precipitation through the  
470 synoptic activity of the African easterly waves (AEW) on its southern flank, along 10°N  
471 [Diedhiou *et al.*, 1999; Fink, 2003]. The main axis of the AEJ is correctly reproduced in the  
472 CTL simulation (Figure 12a). The 4K experiment shows stronger (weaker) easterly winds  
473 south (north) of 10°N, corresponding to a southern displacement of the AEJ, while the  
474 opposite pattern is found in the 4xCO<sub>2</sub> experiment (Figure 12b, c). The modifications in the  
475 AEJ location in the sensitivity experiments are estimated by determining the mean latitudinal  
476 position of the minimum of the zonal wind at 600 hPa in the longitudinal domain [20°W,  
477 30°E]. As a result, the AEJ shifts southward by -3.0° in the 4K MMM, and the multi-model  
478 spread ranges is [-5.3°, -0.2°] (1.5° STD). In the 4xCO<sub>2</sub> experiment, the shift in the MMM is  
479 +1.2°, ranging in [+0.0°, +2.5°] (0.8° STD). The modifications in the AEJ dynamics are  
480 consistent with the precipitation changes displayed in Figure 1 and also a linkage with the  
481 meridional energy profiles in Figure 11 appears evident, being the stronger enthalpy gradient  
482 more favourable to a northward displaced jet in the 4xCO<sub>2</sub> experiment.

483

484 The linkage between the Sahelian precipitation and the thermal forcing over West Africa is  
485 studied by analysing the correlation between the SRI and the meridional gradient of Z925-  
486 700 over West Africa, which is constructed by subtracting Z925-700 averaged north and  
487 south of 20°N in the domain [20°W-30°E, 0°-40°N]. The positive relationship is significant  
488 and strong in ERAI at the interannual time scale ( $r = 0.75$ ), and is captured by CTL in the



489 MMM ( $r = 0.55$ ) and in 8 out of 12 models (Table 2). The relationship is robust both in 4K  
490 and 4xCO<sub>2</sub> (significant in 9 out of 12, and 10 out of 12 models, respectively), with 4xCO<sub>2</sub>  
491 showing higher correlations than 4K in the MMM ( $r = 0.73$  and  $r = 0.52$ , respectively) and in  
492 9 out of 12 models (Table 2).

493

494 The coherence between precipitation and regional monsoonal circulation is studied by  
495 correlating the SRI and the West African Monsoon Index (WAMI), which is defined in the  
496 domain [20°W-20°E, 3°-13°N] as the difference between the standardized wind modulus at  
497 850 hPa and the standardized zonal wind at 200 hPa [Garric *et al.*, 2002]. The WAMI is used  
498 as an indicator of the intensity of the regional monsoonal cell, and is positively correlated to  
499 the Sahel precipitation at the interannual time scale [Fontaine *et al.*, 1995]. GPCP data and  
500 ERAI reanalysis confirm the correlation for the period 1979-2008 ( $r = 0.77$ ), and CTL  
501 simulations partially reproduce the correlation ( $r = 0.46$  in the MMM), with 7 out of 12  
502 models showing significant positive correlations (Table 2). Interestingly, in the 4K  
503 experiment the relationship is even less robust than in CTL ( $r = 0.34$  in the MMM, significant  
504 in 4 out of 12 models), while robustness is recovered in 4xCO<sub>2</sub> ( $r = 0.61$  in the MMM,  
505 significant in 11 out of 12 models), with 10 out of 12 models showing higher correlations  
506 than in 4K (Table 2).

507

508 Finally, the changes in relationship between the mid-troposphere regional circulation and the  
509 Sahelian precipitation are studied by analysing the correlation between SRI and AEJ, which  
510 is represented through the zonal wind at 600 hPa. The positive correlation between  
511 precipitation and intensity and meridional displacement over West Africa of the AEJ is  
512 illustrated in the CTL simulation, showing positive (negative) correlations along the 10°N  
513 (18°N) axis (Figure 13a). When the correlation is computed in the sensitivity experiments, the  
514 general pattern is conserved in both 4K and 4xCO<sub>2</sub>, but significance is strongly reduced in  
515 the former, while the signal is robust in the latter, where the axis over West Africa is also  
516 better represented (Figure 13b, c).

517

518 The analysis of the WAM regional drivers points out the strong impact of the idealized  
519 forcings on the main features of the regional dynamics, namely the SHL, the meridional  
520 energy gradient, and the AEJ. Specifically, the CO<sub>2</sub> concentration increase is particularly  
521 effective in modifying the dynamics over the Sahara desert, favouring the mechanisms  
522 associated with wet conditions in the Sahel. Furthermore, while the SST warming appears to

523 weaken the linkage between precipitation and regional dynamics, the CO<sub>2</sub> increase clearly  
524 strengthens the coupling.

525

## 526 **6. Summary and conclusions**

527

528 In this study, the mechanisms linking the WAM dynamics to the CO<sub>2</sub> forcing are  
529 investigated. To this aim, extreme idealized conditions are imposed in multi-model climate  
530 simulations to maximise and clearly separate the response of the WAM dynamics to the  
531 opposite and competing effects induced by CO<sub>2</sub>. On the one hand, the effect of the direct  
532 radiative forcing, which tends to reinforce the monsoonal circulation, is studied in sensitivity  
533 experiments run with prescribed observed SST and quadrupled CO<sub>2</sub> concentration. On the  
534 other hand, the indirect CO<sub>2</sub> effect mediated by the global SST warming, which weakens the  
535 monsoon, is studied in sensitivity experiments where the global SST is uniformly increased  
536 by 4 K.

537

538 The precipitation response over Sahel shows opposite modifications, with dry conditions  
539 associated with the SST warming (-0.6 mm/day in the Sahel, down to -0.8 mm/day in western  
540 Sahel), while increased CO<sub>2</sub> concentration leads to wet anomalies (+0.5 mm/day in the  
541 Sahel, up to +0.7 mm/day in eastern Sahel). Moreover, the modifications in the precipitation  
542 patterns appear associated with coherent modifications in the monsoonal circulation, namely  
543 in the meridional thermal forcing, the monsoonal flow in the lower troposphere, the AEJ and  
544 TEJ intensity and position, and in the regional meridional overturning circulation.

545

546 The analysis of dynamics aspects reveals that the global SST warming acts at the global  
547 scale, weakening the connection between precipitation and the regional drivers. A warmer  
548 Global Ocean reinforces evaporation in the Tropics, resulting in an inhibition of the  
549 monsoonal flow and the ITCZ migration over West Africa. Conversely, the CO<sub>2</sub> increase  
550 strengthens the coupling between precipitation and the regional dynamics over West Africa.  
551 The local atmospheric radiative forcing, resulting from the direct action of the CO<sub>2</sub> increase,  
552 warms the Sahara and enhances evaporation over the Sahel, producing specific humidity  
553 anomalies in the low troposphere. The consequent increase in the low tropospheric latent heat  
554 and enthalpy content over Sahara and Sahel reinforces the MSE meridional gradient over  
555 West Africa, leading to more intense monsoonal circulation and precipitation. More  
556 specifically, the westerly flow over the Sahel is reinforced, accompanied by a northward shift

557 in moisture convergence (+0.6°) and AEJ (+2.7°). Upward anomalies are seen in the vertical  
558 motions over the Sahel and the Sahara, which are associated with a strengthening of the  
559 meridional overturning circulation, and a 1° northward displacement of the SHL,  
560 respectively. Finally, the anomalous westerly flow in the lower troposphere, and a reinforced  
561 TEJ (+2.9 m/s), connect a coherent zonal circulation cell over West Africa. In this picture, the  
562 modifications in the atmospheric dynamics over the Sahara appear to be particularly  
563 important. A schematic of the WAM dynamics and precipitation in the idealized simulations  
564 is presented in Figure 14.

565

566 The general agreement among models in their response to the idealized conditions  
567 demonstrates the robustness of the identified mechanisms linking the WAM dynamics to the  
568 SST and CO<sub>2</sub> idealized forcings, and reflects the fact that the simulation of these mechanisms  
569 does not depend on the model physics, nor on the model performance. This robustness also  
570 confirms the value of using idealized forcings to maximise and separate the response of the  
571 climate system. Furthermore, the similarity of the future projections in the RCP8.5 scenario  
572 with the combined responses in the 4K and 4xCO<sub>2</sub> simulations, though only qualitative,  
573 clearly shows how the competition between the global SST warming and the CO<sub>2</sub> increase is  
574 a key factor in driving the WAM variability and change. Indeed, it is argued that the large  
575 spread in the simulation of the WAM dynamics and precipitation may be related to an  
576 unbalanced model response to the direct and indirect CO<sub>2</sub> forcings, i.e., dry (wet) biases may  
577 be produced by specific model sensitivity to the SST warming (CO<sub>2</sub> increase). The analysed  
578 models show different sensitivity to the idealized forcings, leading to a sizable spread in the  
579 amplitude of the response. Therefore, the quantitative evaluation of the impact of direct and  
580 indirect CO<sub>2</sub> forcings on the WAM dynamics, and the assessment of the model ability to  
581 reproduce the amplitude of this impact, still need to be addressed. This could be achieved by  
582 analysing model simulations of periods well covered with observations and one of the two  
583 mechanisms is clearly dominant. In this respect, the 20th century and/or the mid-Holocene  
584 could be optimal candidates to analyse the processes responding to changes in the land-sea  
585 contrast, and this will be the subject of a forthcoming study.

586

## 587 **Acknowledgements**

588

589 This work benefitted from the support of the Agence Nationale de la Recherche (ANR) grant  
590 ANR-10-LABX-18-01 of the national Programme Investissements d'Avenir. Funding for this

591 work was also provided by Laboratoire d'excellence Institut Pierre Simon Laplace (L-IPSL).  
592 We acknowledge the World Climate Research Programme's Working Group on Coupled  
593 Modelling, which is responsible for CMIP, and we thank the climate modelling groups (listed  
594 in Table 1 of this paper) for producing and making available their model output. For CMIP  
595 the U.S. Department of Energy's Program for Climate Model Diagnosis and Intercomparison  
596 provides coordinating support and led development of software infrastructure in partnership  
597 with the Global Organization for Earth System Science Portals. Authors thank A. Evan for  
598 useful discussions, and two anonymous reviewers for their insightful comments that have  
599 greatly improved the quality of the paper.

600

## 601 **References**

- 602 Adler, R. F. et al. (2003), The Version-2 Global Precipitation Climatology Project (GPCP)  
603 Monthly Precipitation Analysis (1979–Present), *J. Hydrometeorol.*, *4*, 1147–1167,  
604 doi:10.1175/1525-7541(2003)004<1147:TVGPCP>2.0.CO;2.
- 605 Bader, J., and M. Latif (2003), The impact of decadal-scale Indian Ocean sea surface  
606 temperature anomalies on Sahelian rainfall and the North Atlantic Oscillation, *Geophys.*  
607 *Res. Lett.*, *30*, 1–4, doi:10.1029/2003GL018426.
- 608 Biasutti, M. (2013), Forced Sahel rainfall trends in the CMIP5 archive, *J. Geophys. Res.*  
609 *Atmos.*, *118*, 1613–1623, doi:10.1002/jgrd.50206.
- 610 Biasutti, M., A. H. Sobel, and S. J. Camargo (2009), The Role of the Sahara Low in  
611 Summertime Sahel Rainfall Variability and Change in the CMIP3 Models, *J. Clim.*, *22*,  
612 5755–5771, doi:10.1175/2009JCLI2969.1.
- 613 Bony, S., G. Bellon, D. Klocke, S. Sherwood, S. Fermepin, and S. Denvil (2013), Robust  
614 direct effect of carbon dioxide on tropical circulation and regional precipitation, *Nat.*  
615 *Geosci.*, *6*, 447–451, doi:10.1038/ngeo1799.
- 616 Caminade, C., and L. Terray (2010), Twentieth century sahel rainfall variability as simulated  
617 by the ARPEGE AGCM, and future changes, *Clim. Dyn.*, *35*, 75–94,  
618 doi:10.1007/s00382-009-0545-4.
- 619 Chauvin, F., R. Roehrig, and J. P. Lafore (2010), Intraseasonal variability of the saharan heat  
620 low and its link with midlatitudes, *J. Clim.*, *23*, 2544–2561,  
621 doi:10.1175/2010JCLI3093.1.
- 622 Chen, T. C. (2005), Maintenance of the midtropospheric North African summer circulation;  
623 Saharan high and African easterly jet, *J. Clim.*, *18*, 2943–2962, doi:10.1175/JCLI3446.1.
- 624 Cook, K. H. (1999), Generation of the African easterly jet and its role in determining West  
625 African precipitation, *J. Clim.*, *12*, 1165–1184, doi:10.1175/1520-  
626 0442(1999)012<1165:GOTAEJ>2.0.CO;2.
- 627 Cook, K. H., and E. K. Vizy (2006), Coupled model simulations of the West African  
628 monsoon system: Twentieth- and twenty-first-century simulations, *J. Clim.*, *19*, 3681–  
629 3703, doi:10.1175/JCLI3814.1.
- 630 Diedhiou, a., S. Janicot, a. Viltard, P. De Felice, and H. Laurent (1999), Easterly wave

631 regimes and associated convection over West Africa and tropical Atlantic: Results from  
632 the NCEP/NCAR and ECMWF reanalyses, *Clim. Dyn.*, *15*, 795–822,  
633 doi:10.1007/s003820050316.

634 Evan, A. T., C. Flamant, C. Lavaysse, C. Kocha, and A. Saci (2015), Water Vapor–Forced  
635 Greenhouse Warming over the Sahara Desert and the Recent Recovery from the  
636 Sahelian Drought, *J. Clim.*, *28*, 108–123, doi:10.1175/JCLI-D-14-00039.1.

637 Fink, A. H. (2003), Spatiotemporal variability of the relation between African Easterly  
638 Waves and West African Squall Lines in 1998 and 1999, *J. Geophys. Res.*, *108*, 1–17,  
639 doi:10.1029/2002JD002816.

640 Fontaine, B., and N. Philippon (2000), Seasonal Evolution of Boundary Layer Heat Content  
641 in the West African Monsoon From the Ncep / Ncar, *Int. J. Climatol.*, *20*, 1777–1790.

642 Fontaine, B., S. Janicot, and V. Moron (1995), Rainfall anomaly patterns and wind field  
643 signals over West Africa in August (1958-1989), *J. Clim.*, *8*, 1503–1510.

644 Fontaine, B., J. Garcia-Serrano, P. Roucou, B. Rodriguez-Fonseca, T. Losada, F. Chauvin, S.  
645 Gervois, S. Sijikumar, P. Ruti, and S. Janicot (2010), Impacts of warm and cold  
646 situations in the Mediterranean basins on the West African monsoon: Observed  
647 connection patterns (1979-2006) and climate simulations, *Clim. Dyn.*, *35*, 95–114,  
648 doi:10.1007/s00382-009-0599-3.

649 Fontaine, B., M. Gaetani, A. Ullmann, and P. Roucou (2011), Time evolution of observed  
650 July-September sea surface temperature-Sahel climate teleconnection with removed  
651 quasi-global effect (1900-2008), *J. Geophys. Res. Atmos.*, *116*, 1–17,  
652 doi:10.1029/2010JD014843.

653 Gaetani, M., B. Fontaine, P. Roucou, and M. Baldi (2010), Influence of the Mediterranean  
654 Sea on the West African monsoon: Intraseasonal variability in numerical simulations, *J.*  
655 *Geophys. Res.*, *115*, 1–17, doi:10.1029/2010JD014436.

656 Garric, G., H. Douville, and M. Déqué (2002), Prospects for improved seasonal predictions of  
657 monsoon precipitation over Sahel, *Int. J. Climatol.*, *22*, 331–345, doi:10.1002/joc.736.

658 Giannini, A. (2010), Mechanisms of Climate Change in the Semiarid African Sahel: The  
659 Local View, *J. Clim.*, *23*, 743–756, doi:10.1175/2009JCLI3123.1.

660 Giannini, A., R. Saravanan, and P. Chang (2003), Oceanic forcing of Sahel rainfall on  
661 interannual to interdecadal time scales, *Science*, *302*, 1027–1030,  
662 doi:10.1126/science.1089357.

663 Haarsma, R. J., F. M. Selten, S. L. Weber, and M. Kliphuis (2005), Sahel rainfall variability  
664 and response to greenhouse warming, *Geophys. Res. Lett.*, *32*, 1–4,  
665 doi:10.1029/2005GL023232.

666 Held, I. M., T. L. Delworth, J. Lu, K. L. Findell, and T. R. Knutson (2005), Simulation of  
667 Sahel drought in the 20th and 21st centuries., *Proc. Natl. Acad. Sci. U. S. A.*, *102*,  
668 17891–17896, doi:10.1073/pnas.0509057102.

669 Hoerling, M., J. Hurrell, J. Eischeid, and A. Phillips (2006), Detection and attribution of  
670 twentieth-century northern and southern African rainfall change, *J. Clim.*, *19*, 3989–  
671 4008, doi:10.1175/JCLI3842.1.

672 IPCC, (2014), Climate Change 2014: Synthesis Report. Contribution of Working Groups I, II  
673 and III to the Fifth Assessment Report of the Intergovernmental Panel on Climate  
674 Change [Core Writing Team, R.K. Pachauri and L.A. Meyer (eds.)]. IPCC, Geneva,  
675 Switzerland, 151 pp.

676 Issa Lélé, M., and P. J. Lamb (2010), Variability of the Intertropical Front (ITF) and Rainfall  
677 over the West African Sudan–Sahel Zone, *J. Clim.*, *23*, 3984–4004,  
678 doi:10.1175/2010JCLI3277.1.

679 Kandji, S. T., L. Verchot, and J. Mackensen (2006), *Climate change and variability in the*  
680 *Sahel region : impacts and adaptation strategies in the agricultural sector*. UNEP &  
681 ICRAF.

682 Lavaysse, C., C. Flamant, S. Janicot, D. J. Parker, J. P. Lafore, B. Sultan, and J. Pelon (2009),  
683 Seasonal evolution of the West African heat low: A climatological perspective, *Clim.*  
684 *Dyn.*, *33*, 313–330, doi:10.1007/s00382-009-0553-4.

685 Lavaysse, C., C. Flamant, S. Janicot, and P. Knippertz (2010a), Links between African  
686 easterly waves, midlatitude circulation and intraseasonal pulsations of the West African  
687 heat low, *Q. J. R. Meteorol. Soc.*, *136*, 141–158, doi:10.1002/qj.555.

688 Lavaysse, C., C. Flamant, and S. Janicot (2010b), Regional-scale convection patterns during  
689 strong and weak phases of the Saharan heat low, *Atmos. Sci. Lett.*, *11*, 255–264,  
690 doi:10.1002/asl.284.

691 Lavaysse, C., J. P. Chaboureau, and C. Flamant (2011), Dust impact on the west african heat  
692 low in summertime, *Q. J. R. Meteorol. Soc.*, *137*, 1227–1240, doi:10.1002/qj.844.

693 Losada, T., B. Rodríguez-Fonseca, S. Janicot, S. Gervois, F. Chauvin, and P. Ruti (2010), A  
694 multi-model approach to the Atlantic Equatorial mode: Impact on the West African  
695 monsoon, *Clim. Dyn.*, *35*, 29–43, doi:10.1007/s00382-009-0625-5.

696 Lu, J., and T. L. Delworth (2005), Oceanic forcing of the late 20th century Sahel drought,  
697 *Geophys. Res. Lett.*, *32*, L22706, doi:10.1029/2005GL023316.

698 Meehl, G. A., C. Covey, K. E. Taylor, T. Delworth, R. J. Stouffer, M. Latif, B. McAvaney,  
699 and J. F. B. Mitchell (2007), THE WCRP CMIP3 Multimodel Dataset: A New Era in  
700 Climate Change Research, *Bull. Am. Meteorol. Soc.*, *88*, 1383–1394,  
701 doi:10.1175/BAMS-88-9-1383.

702 Mohino, E., S. Janicot, and J. Bader (2011), Sahel rainfall and decadal to multi-decadal sea  
703 surface temperature variability, *Clim. Dyn.*, *37*, 419–440, doi:10.1007/s00382-010-0867-  
704 2.

705 Monerie, P. A., B. Fontaine, and P. Roucou (2012), Expected future changes in the African  
706 monsoon between 2030 and 2070 using some CMIP3 and CMIP5 models under a  
707 medium-low RCP scenario, *J. Geophys. Res. Atmos.*, *117*, 1–12,  
708 doi:10.1029/2012JD017510.

709 Monerie, P.-A., P. Roucou, and B. Fontaine (2013), Mid-century effects of Climate Change  
710 on African monsoon dynamics using the A1B emission scenario, *Int. J. Climatol.*, *33*,  
711 881–896, doi:10.1002/joc.3476.

712 Nicholson, S. E. (2013), The West African Sahel: A Review of Recent Studies on the  
713 Rainfall Regime and Its Interannual Variability, *ISRN Meteorol.*, *2013*, 32,  
714 doi:10.1155/2013/453521.

715 Nicholson, S. E. et al. (2003), Validation of TRMM and Other Rainfall Estimates with a  
716 High-Density Gauge Dataset for West Africa. Part I: Validation of GPCP Rainfall  
717 Product and Pre-TRMM Satellite and Blended Products, *J. Appl. Meteorol.*, *42*, 1337–  
718 1354, doi:10.1175/1520-0450(2003)042<1337:VOTAOR>2.0.CO;2.

719 Panthou, G., Vischel, T. and Lebel, T. (2014), Recent trends in the regime of extreme rainfall  
720 in the Central Sahel. *Int. J. Climatol.*, *34*: 3998–4006. doi:10.1002/joc.3984.

- 721 Riahi, K., S. Rao, V. Krey, C. Cho, V. Chirkov, G. Fischer, G. Kindermann, N. Nakicenovic,  
722 and P. Rafaj (2011), RCP 8.5-A scenario of comparatively high greenhouse gas  
723 emissions, *Clim. Change*, *109*, 33–57, doi:10.1007/s10584-011-0149-y.
- 724 Rodríguez-Fonseca, B. et al. (2015), Variability and Predictability of West African Droughts:  
725 A Review on the Role of Sea Surface Temperature Anomalies, *J. Clim.*, *28*, 4034–4060,  
726 doi:10.1175/JCLI-D-14-00130.1.
- 727 Roehrig, R., D. Bouniol, F. Guichard, F. D. Hourdin, and J. L. Redelsperger (2013), The  
728 present and future of the west african monsoon: A process-oriented assessment of  
729 CMIP5 simulations along the AMMA transect, *J. Clim.*, *26*, 6471–6505,  
730 doi:10.1175/JCLI-D-12-00505.1.
- 731 Rowell, D. P. (2001), Teleconnections between the tropical Pacific and the Sahel, *Q. J. R.*  
732 *Meteorol. Soc.*, *127*, 1683–1706, doi:10.1002/qj.49712757512.
- 733 Rowell, D. P. (2013), Simulating SST teleconnections to Africa: What is the state of the art?,  
734 *J. Clim.*, *26*, 5397–5418, doi:10.1175/JCLI-D-12-00761.1.
- 735 Santer, B. D. et al. (2009), Incorporating model quality information in climate change  
736 detection and attribution studies., *Proc. Natl. Acad. Sci. U. S. A.*, *106*, 14778–14783,  
737 doi:10.1073/pnas.0901736106.
- 738 Skinner, C. B., M. Ashfaq, and N. S. Diffenbaugh (2012), Influence of Twenty-First-Century  
739 Atmospheric and Sea Surface Temperature Forcing on West African Climate, *J. Clim.*,  
740 *25*, 527–542, doi:10.1175/2011JCLI4183.1.
- 741 Tanaka, H. L., N. Ishizaki, and A. Kitoh (2004), Trend and interannual variability of Walker,  
742 monsoon and Hadley circulations defined by velocity potential in the upper troposphere,  
743 *Tellus, Ser. A Dyn. Meteorol. Oceanogr.*, *56*, 250–269, doi:10.1111/j.1600-  
744 0870.2004.00049.x.
- 745 Taylor, K. E., R. J. Stouffer, and G. a. Meehl (2012), An overview of CMIP5 and the  
746 experiment design, *Bull. Am. Meteorol. Soc.*, *93*, 485–498, doi:10.1175/BAMS-D-11-  
747 00094.1.
- 748 Thorncroft, C. D., and M. Blackburn (1999), Maintenance of the African easterly jet, *Q. J. R.*  
749 *Meteorol. Soc.*, *125*, 763–786, doi:10.1002/qj.49712555502.
- 750 Ting, M., Y. Kushnir, R. Seager, and C. Li (2009), Forced and Internal Twentieth-Century  
751 SST Trends in the North Atlantic, *J. Clim.*, *22*, 1469–1481,  
752 doi:10.1175/2008JCLI2561.1.
- 753 Trenberth, K.E., P.D. Jones, P. Ambenje, R. Bojariu, D. Easterling, A. Klein Tank, D. Parker,  
754 F. Rahimzadeh, J.A. Renwick, M. Rusticucci, B. Soden and P. Zhai, 2007:  
755 Observations: Surface and Atmospheric Climate Change. In: *Climate Change 2007: The*  
756 *Physical Science Basis. Contribution of Working Group I to the Fourth Assessment*  
757 *Report of the Intergovernmental Panel on Climate Change [Solomon, S., D. Qin, M.*  
758 *Manning, Z. Chen, M. Marquis, K.B. Averyt, M. Tignor and H.L. Miller (eds.)].*  
759 *Cambridge University Press, Cambridge, United Kingdom and New York, NY, USA.*
- 760 Villamayor, J., and E. Mohino (2015), Robust Sahel drought due to the Interdecadal Pacific  
761 Oscillation in CMIP5 simulations, *Geophys. Res. Lett.*, *42*, 1214–1222,  
762 doi:10.1002/2014GL062473.
- 763 Vizzy, E. K., and K. H. Cook (2009), A mechanism for African monsoon breaks:  
764 Mediterranean cold air surges, *J. Geophys. Res. Atmos.*, *114*, 1–19,  
765 doi:10.1029/2008JD010654.

- 766 Xue, Y. et al. (2010), Intercomparison and analyses of the climatology of the West African  
767 Monsoon in the West African Monsoon Modeling and Evaluation project (WAMME)  
768 first model intercomparison experiment, *Clim. Dyn.*, 35, 3–27, doi:10.1007/s00382-010-  
769 0778-2.
- 770 Zhang, R., and T. L. Delworth (2006), Impact of Atlantic multidecadal oscillations on  
771 India/Sahel rainfall and Atlantic hurricanes, *Geophys. Res. Lett.*, 33, L17712,  
772 doi:10.1029/2006GL026267.  
773



774 **Tables**

775

776 Table 1: Models analysed. CMIP5 model information and outputs are available through the

777 Earth System Grid Federation archive (<http://cmip-pcmdi.llnl.gov/cmip5>).

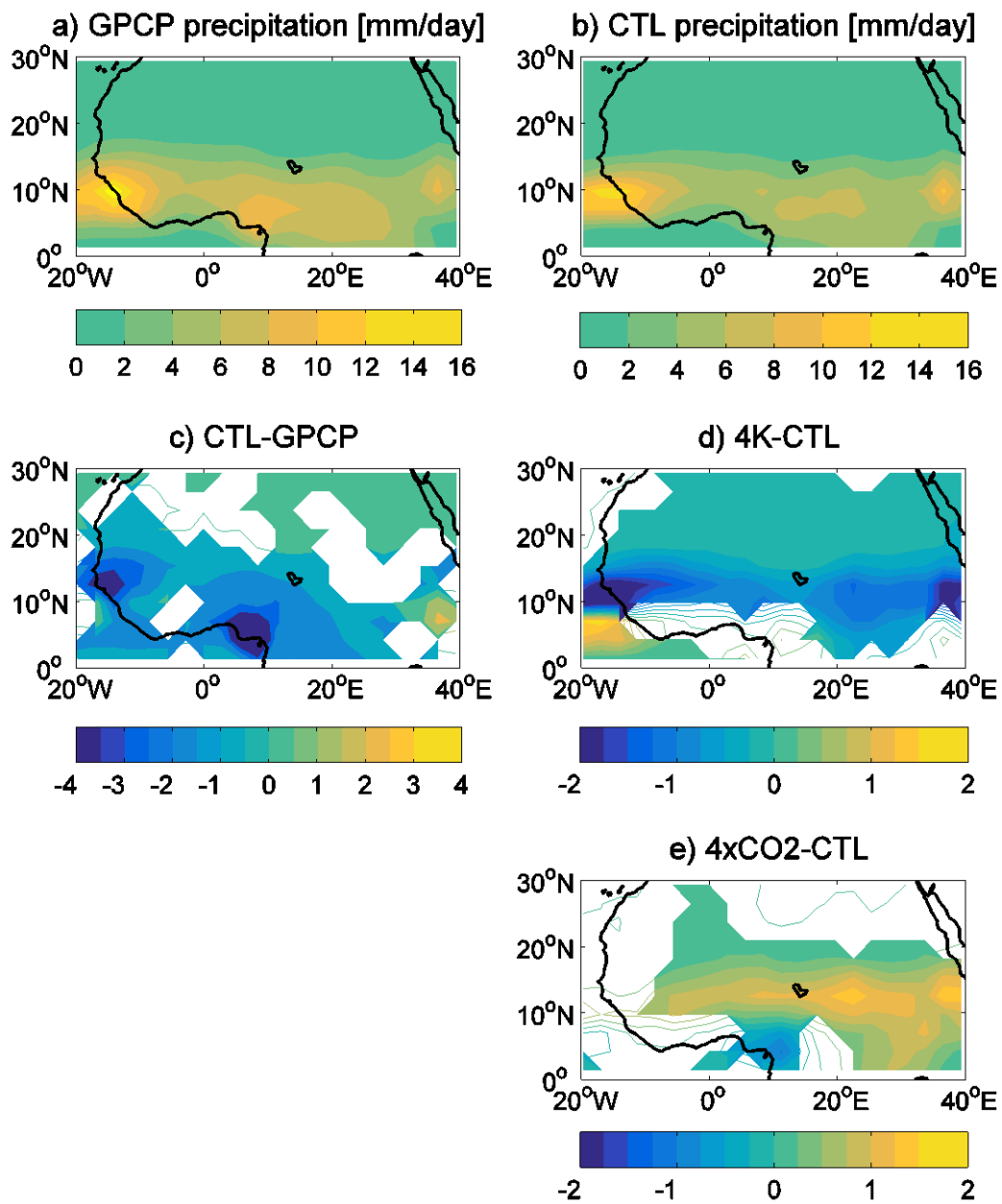
| Country           | Modelling centre   | Model        | Resolution (lat-lon) |
|-------------------|--|--------------|----------------------|
| China             | Beijing Climate Center, China<br>Meteorological Administration   | BCC-CSM1-1   | T42 (~2.8°)          |
| Canada            | Canadian Centre for Climate Modelling<br>and Analysis  | CanAM4       | T42 (~2.8°)          |
| USA               | National Centre for Atmospheric<br>Research  | CCSM4        | 1.25° x 0.9°         |
| France            | Centre National de Recherches<br>Météorologiques / Centre Européen de<br>Recherche et Formation Avancée en<br>Calcul Scientifique  | CNRM-CM5     | T127 (~1.4°)         |
| China             | LASG, Institute of Atmospheric<br>Physics, Chinese Academy of Sciences<br>and CESS, Tsinghua University  | FGOALS-g2    | 3.0° x 2.8°          |
| United<br>Kingdom | Met Office Hadley Centre   | HadGEM2-A    | 1.25° x 1.875°       |
| France            | Institut Pierre-Simon Laplace  | IPSL-CM5A-LR | 1.875° x 3.75°       |
|                   |  | IPSL-CM5B-LR | 1.875° x 3.75°       |
| Japan             | Atmosphere and Ocean Research<br>Institute (The University of Tokyo),<br>National Institute for Environmental<br>Studies, and Japan Agency for Marine-<br>Earth Science and Technology | MIROC5       | T127 (~1.4°)         |
| Germany           | Max-Planck-Institut für Meteorologie<br>(Max Planck Institute for Meteorology)   | MPI-ESM-LR   | T63 (~1.9°)          |
|                   |  | MPI-ESM-MR   | T63 (~1.9°)          |
| Japan             | Meteorological Research Institute  | MRI-CGCM3    | T159 (~1.125°)       |

778

779 Table 2: Correlation coefficients between the 1979-2008 time series of the Z925-700  
780 meridional gradient (GradZ) and the West African Monsoon Index (WAMI), and the  
781 Sahelian rainfall index (SRI), in three experiments and in observations and reanalysis. The  
782 linear trend has been removed from the time series. In bold, positive correlations significant  
783 at 95% confidence level. The highest value between 4K and 4xCO2 is underlined.

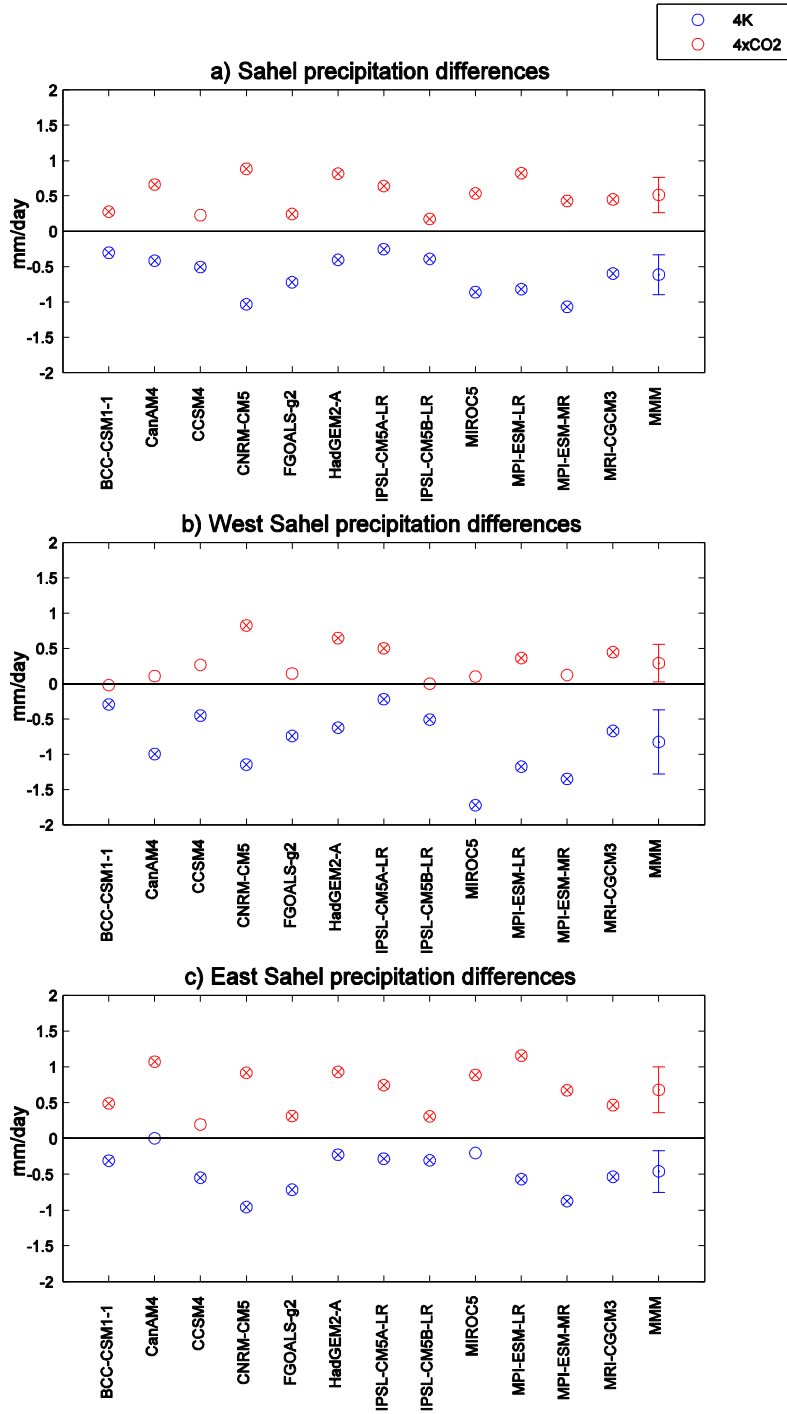
|              | GradZ       |                    |                    | WAMI        |                    |                    |
|--------------|-------------|--------------------|--------------------|-------------|--------------------|--------------------|
|              | CTL         | 4K                 | 4xCO2              | CTL         | 4K                 | 4xCO2              |
| BCC-CSM1-1   | 0.07        | <b>0.38</b>        | <u><b>0.63</b></u> | 0.06        | -0.50              | <u><b>0.67</b></u> |
| CanAM4       | 0.29        | <b>0.48</b>        | <u><b>0.64</b></u> | 0.21        | 0.21               | <u><b>0.83</b></u> |
| CCSM4        | <b>0.45</b> | <b>0.57</b>        | <u><b>0.66</b></u> | <b>0.52</b> | <b>0.66</b>        | <u><b>0.69</b></u> |
| CNRM-CM5     | <b>0.47</b> | 0.33               | <u><b>0.71</b></u> | <b>0.37</b> | <b>0.38</b>        | <u><b>0.58</b></u> |
| FGOALS-g2    | <b>0.38</b> | <b>0.60</b>        | <u><b>0.70</b></u> | -0.03       | -0.71              | <u><b>0.50</b></u> |
| HadGEM2-A    | <b>0.68</b> | <b>0.54</b>        | <u><b>0.81</b></u> | <b>0.53</b> | 0.09               | <u><b>0.65</b></u> |
| IPSL-CM5A-LR | <b>0.43</b> | <u><b>0.54</b></u> | 0.20               | <b>0.48</b> | <u><b>0.38</b></u> | 0.25               |
| IPSL-CM5B-LR | -0.20       | <u>0.23</u>        | 0.15               | -0.03       | -0.71              | <u><b>0.50</b></u> |
| MIROC5       | 0.30        | <u><b>0.78</b></u> | <b>0.57</b>        | <b>0.50</b> | <u><b>0.69</b></u> | <b>0.60</b>        |
| MPI-ESM-LR   | <b>0.39</b> | <b>0.40</b>        | <u><b>0.71</b></u> | <b>0.54</b> | 0.08               | <u><b>0.76</b></u> |
| MPI-ESM-MR   | <b>0.59</b> | 0.06               | <u><b>0.66</b></u> | <b>0.56</b> | -0.22              | <u><b>0.54</b></u> |
| MRI-CGCM3    | <b>0.69</b> | <b>0.44</b>        | <u><b>0.58</b></u> | 0.20        | 0.24               | <u><b>0.41</b></u> |
| MMM          | <b>0.55</b> | <b>0.52</b>        | <u><b>0.73</b></u> | <b>0.46</b> | 0.34               | <u><b>0.61</b></u> |
| ERA1, GPCP   | <b>0.75</b> |                    |                    | <b>0.77</b> |                    |                    |

784



786

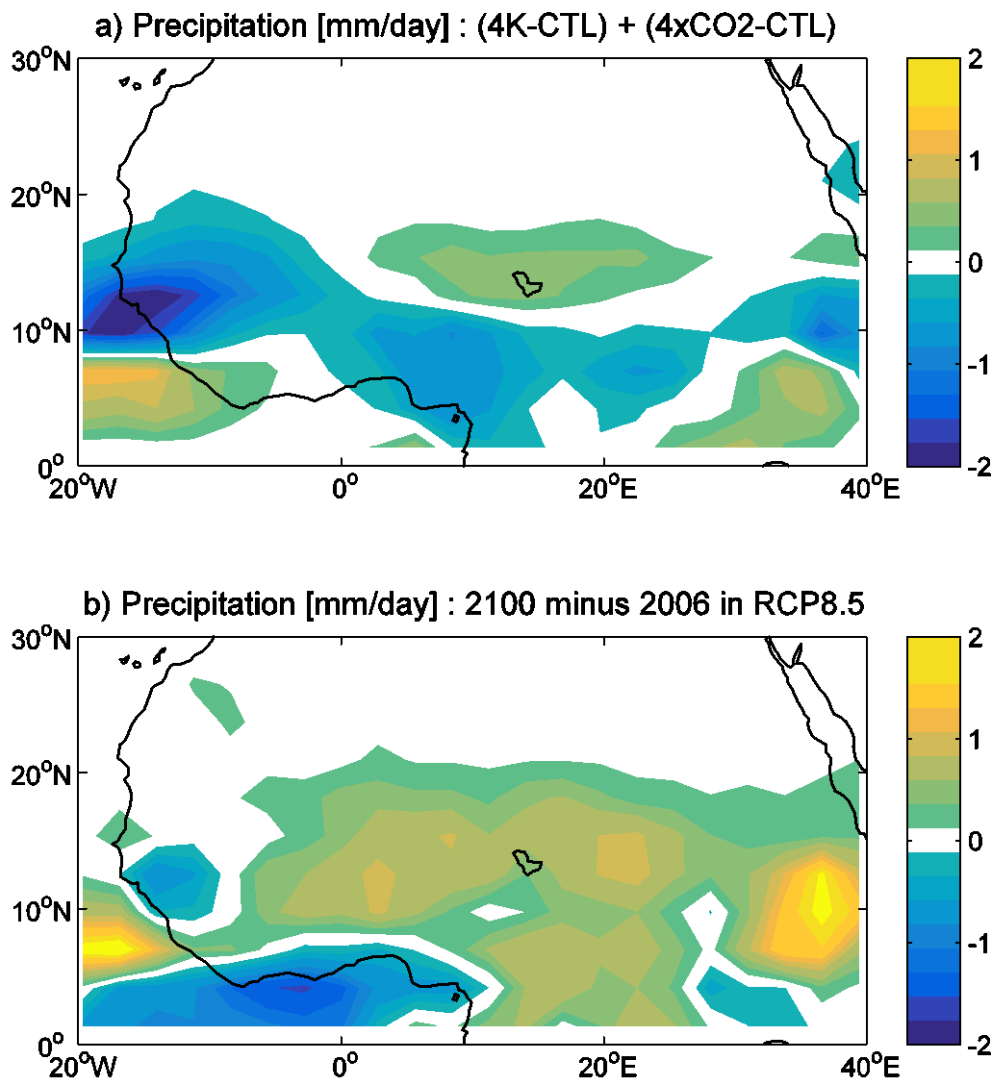
787 Figure 1: JAS precipitation [mm/day]: (a) GPCP observations, (b) MMM of the CTL  
 788 simulation and (c) difference with observations, and MMM differences between (d) 4K and  
 789 (e) 4xCO<sub>2</sub> experiments and CTL simulation. Significant differences at 95% confidence level  
 790 are shaded. Note that the reference scale is not the same for the climatology, the model bias,  
 791 and the sensitivity experiment responses.



792

793 Figure 2: JAS precipitation [mm/day]: differences between the 4K (blue circles) and 4xCO<sub>2</sub>  
 794 (red circles) experiments and the CTL simulation, in (a) Sahel [10°-20°N, 20°W-30°E], (b)  
 795 western (west of 0°E) and (c) eastern Sahel (east of 0°E). Vertical bars for the MMM indicate  
 796 multi-model STD, crosses indicate significant differences at 95% confidence level.

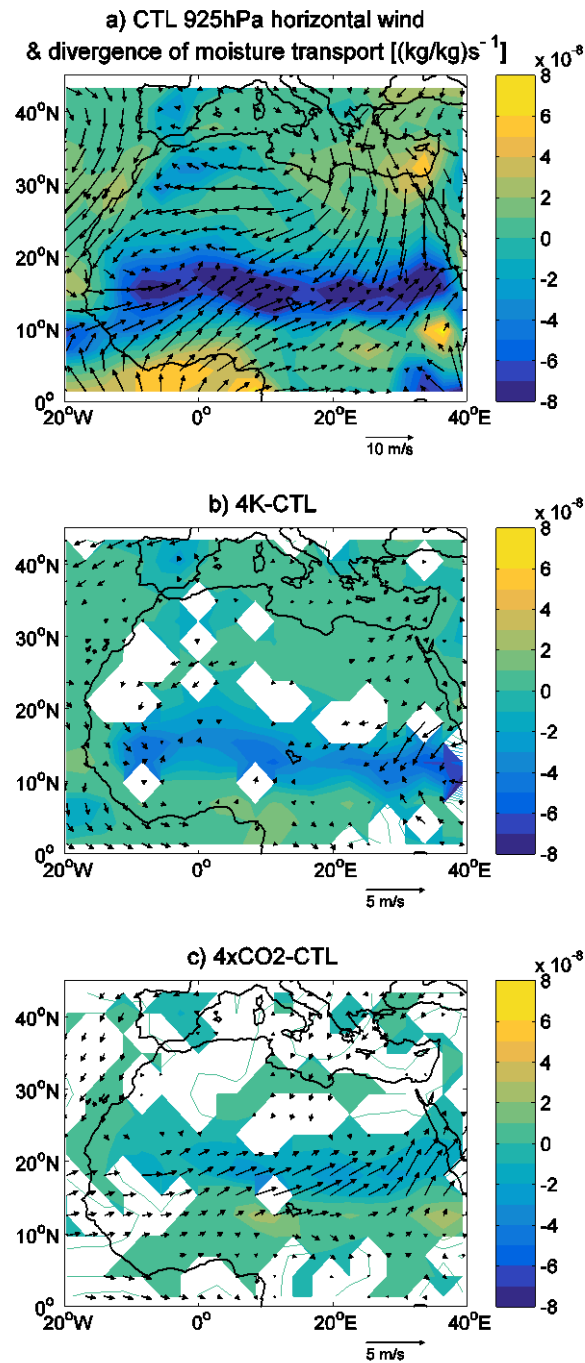
797



798

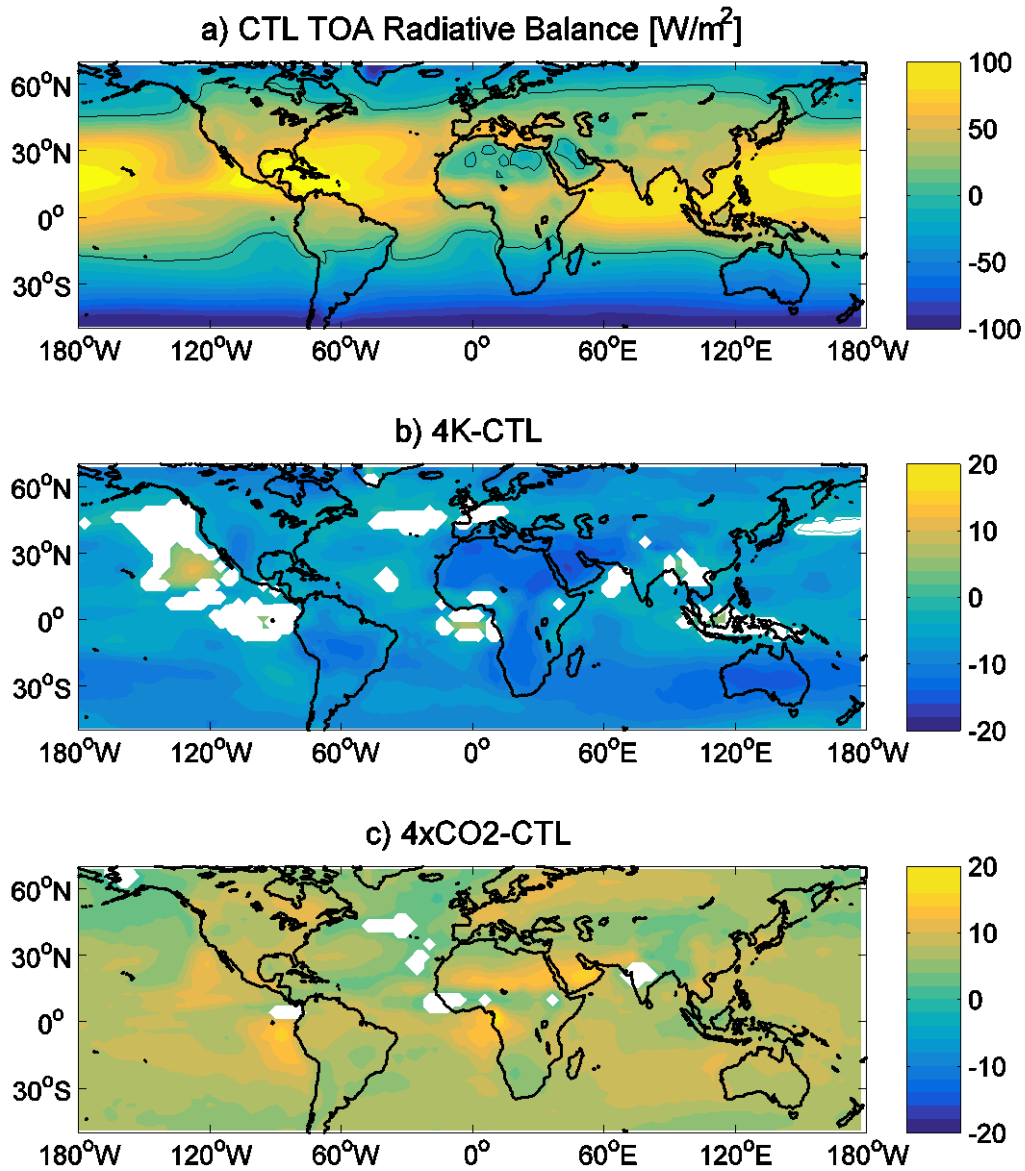
799 Figure 3: MMM of the JAS precipitation [mm/day]: (a) combination of the differences  
 800 between 4K and 4xCO2 experiments and CTL simulation, computed as the sum of the  
 801 responses displayed in Figure 1; and (b) 21st century changes in the RCP8.5 scenario,  
 802 computed as the year 2100 minus 2006 difference assuming a linear trend over the period  
 803 2006-2100.

804



805

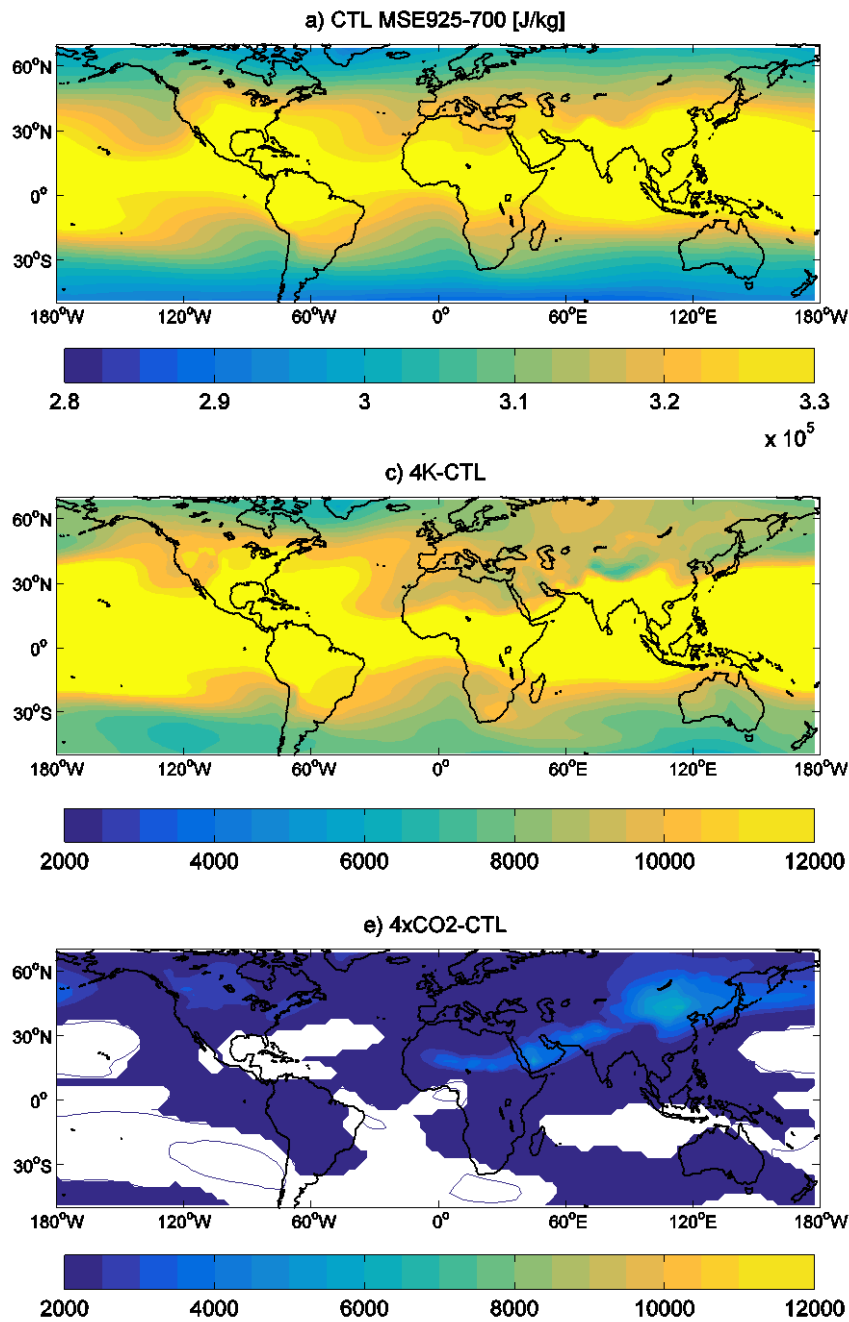
806 Figure 4: JAS horizontal wind (vectors) and moisture transport divergence (contour and  
 807 shadings) at 925 hPa  $[\text{kg}/\text{kg}\cdot\text{s}^{-1}]$ : (a) MMM of the CTL simulation and MMM differences  
 808 between (b) 4K and (c) 4xCO<sub>2</sub> experiments and CTL simulation. Vectors and shadings  
 809 indicate differences significant at 95% confidence level. Note that the reference vector is not  
 810 the same for the CTL simulation and the sensitivity experiment responses.



811

812 Figure 5: JAS global net radiative balance, defined at the top of the atmosphere as the  
 813 difference between the downward shortwave radiation and the sum of upward shortwave and  
 814 longwave radiation: (a) MMM of the CTL simulation and MMM differences between (b) 4K  
 815 and (c) 4xCO<sub>2</sub> experiments and CTL simulation. In panel (a), black contours highlight the  
 816 separation between positive and negative values. Significant differences at 95% confidence  
 817 level are shaded. Note that the reference scale is not the same for the CTL simulation and the  
 818 sensitivity experiment responses.

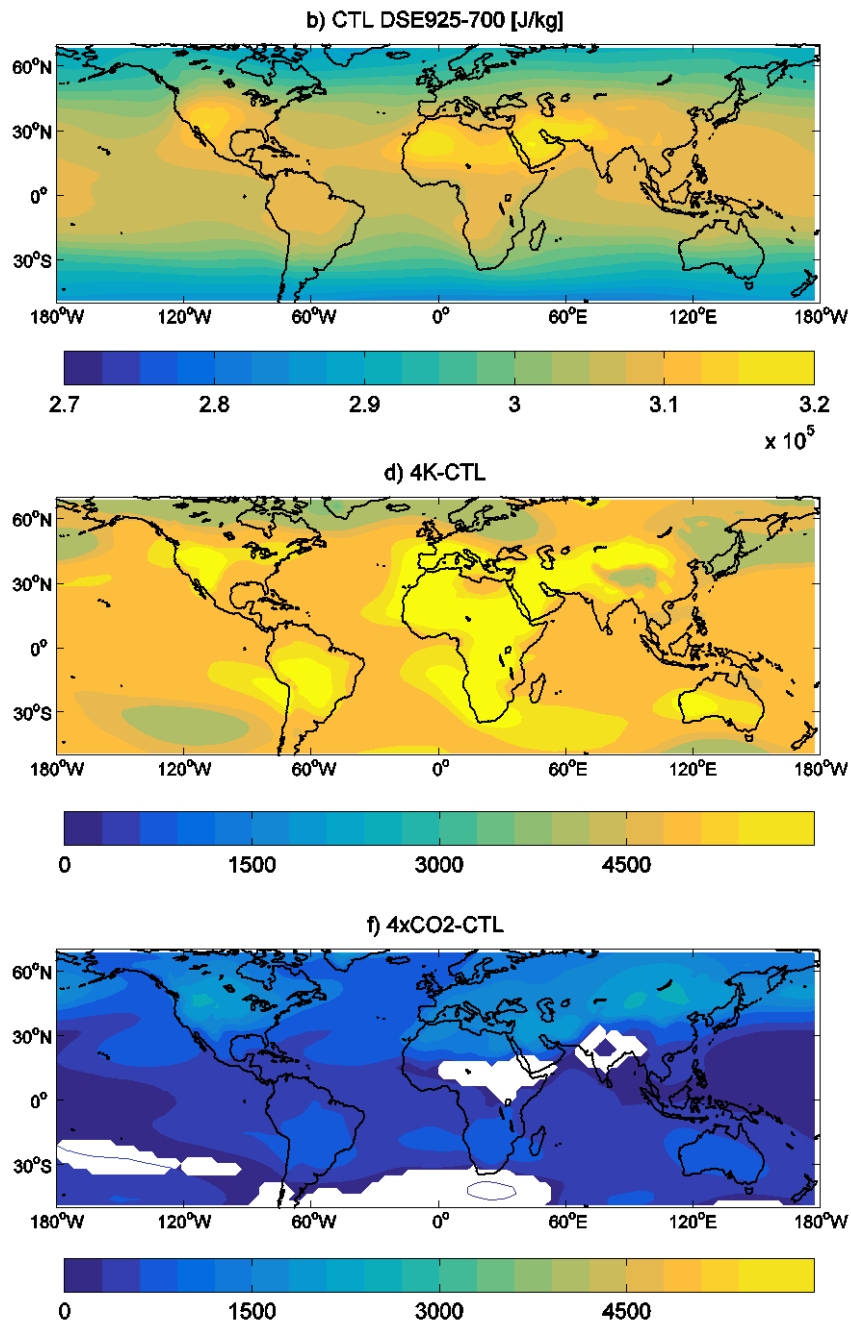
819



820

821 Figure 6: JAS (left) MSE and (right) DSE integrated in the 925-700 hPa layer: (top) MMM of  
 822 the CTL simulation and MMM differences between (middle) 4K and (bottom) 4xCO<sub>2</sub>  
 823 experiments and CTL simulation. Significant differences at 95% confidence level are shaded.  
 824 Note that the reference scale is not the same for the CTL simulation and the sensitivity  
 825 experiment responses.

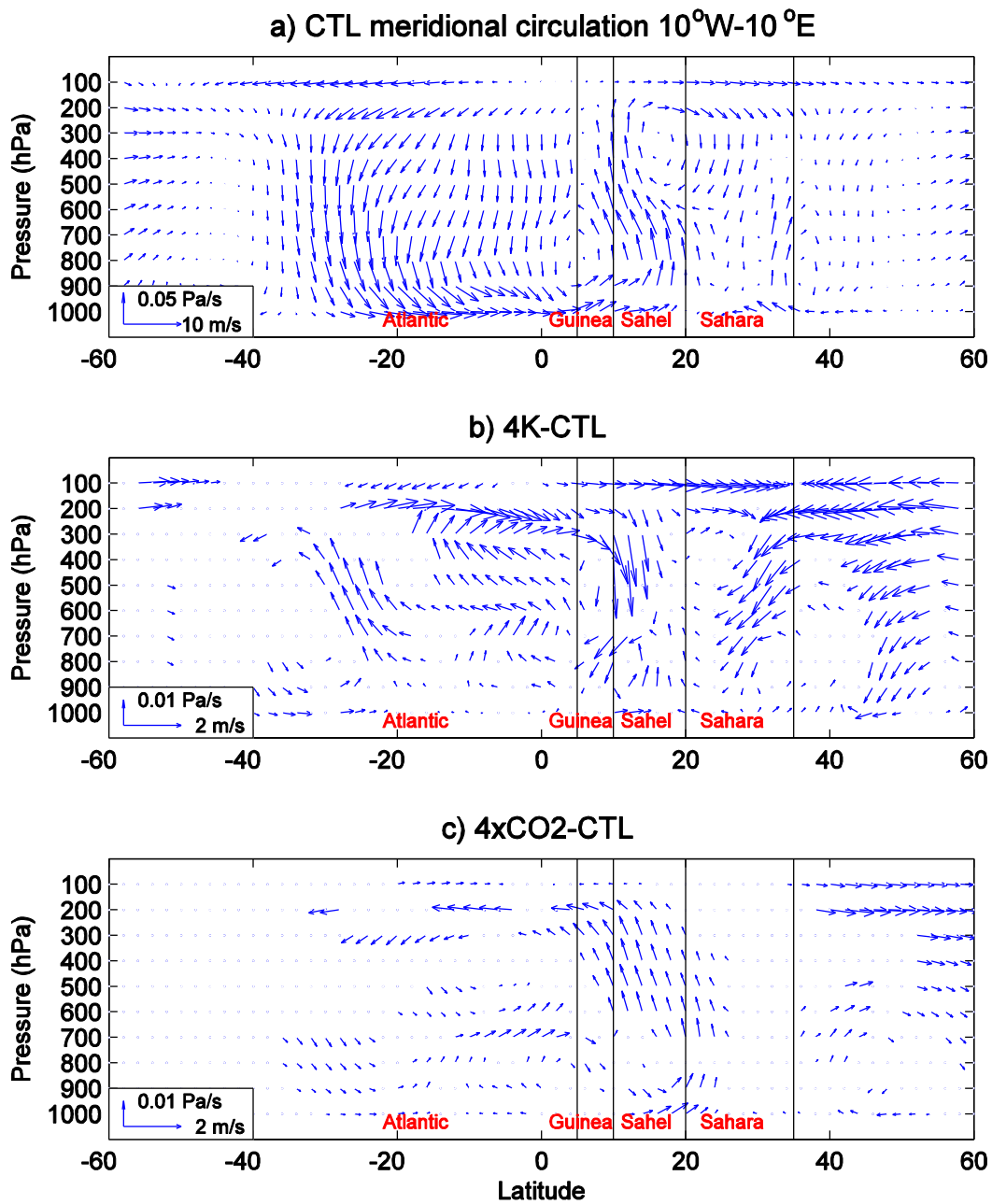




826

827 Figure 6: Continued.

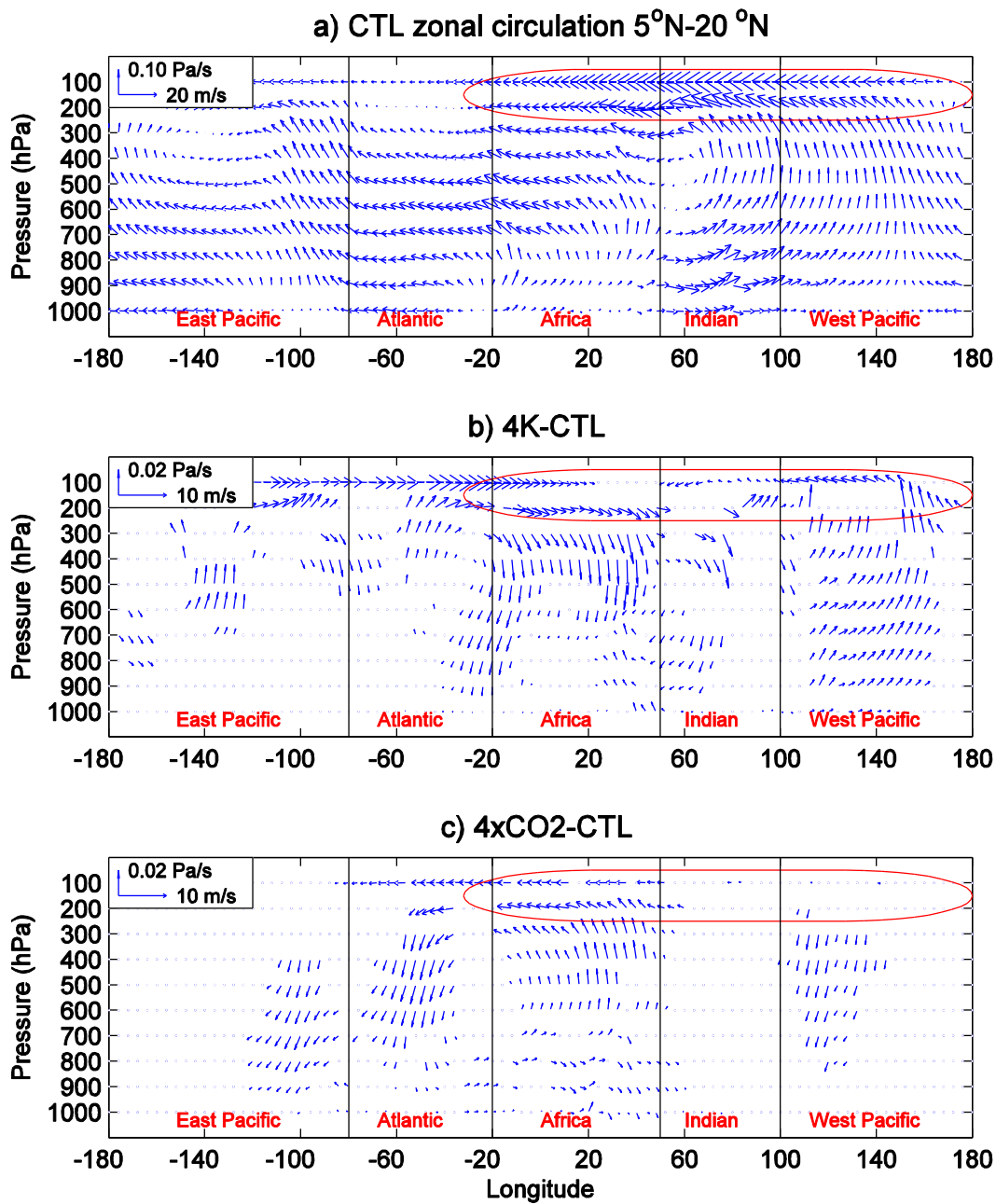
828



829

830 Figure 7: JAS meridional circulation averaged in [10°W, 10°E], displayed through meridional  
 831 wind [m/s] and vertical velocity [Pa/s] vectors in latitude-pressure plots. (a) MMM of the  
 832 CTL simulation and MMM differences between (b) 4K and (c) 4xCO2 experiments and CTL  
 833 simulations. Significant differences at 95% confidence level are displayed. Note that the  
 834 reference vectors are not the same for the CTL simulation and the sensitivity experiment  
 835 responses.

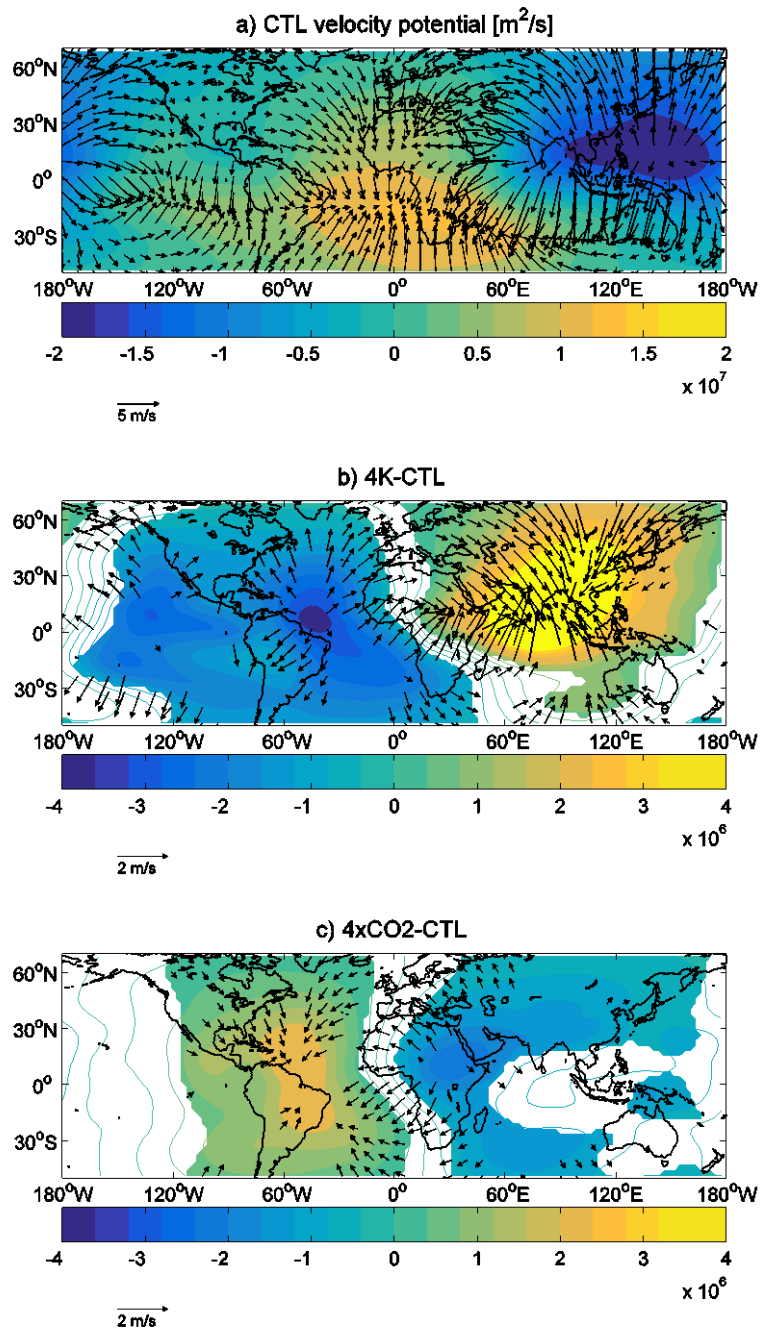
836



837

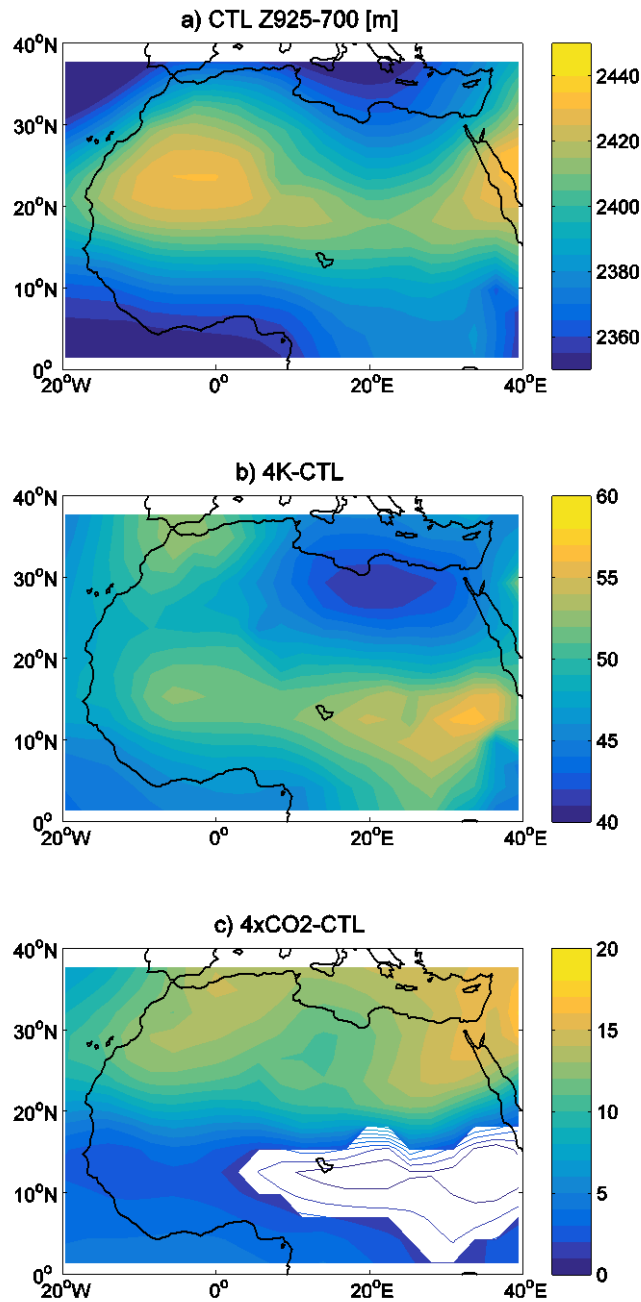
838 Figure 8: JAS zonal circulation averaged in [5°N, 20°N], displayed through zonal wind [m/s]  
 839 and vertical velocity [Pa/s] vectors in latitude-pressure plots. (a) MMM of the CTL  
 840 simulation and MMM differences between (b) 4K and (c) 4xCO<sub>2</sub> experiments and CTL  
 841 simulation. The TEJ region in the upper troposphere is highlighted in red. Significant  
 842 differences at 95% confidence level are displayed. Note that the reference vectors are not the  
 843 same for the CTL simulation and the sensitivity experiment responses.

844



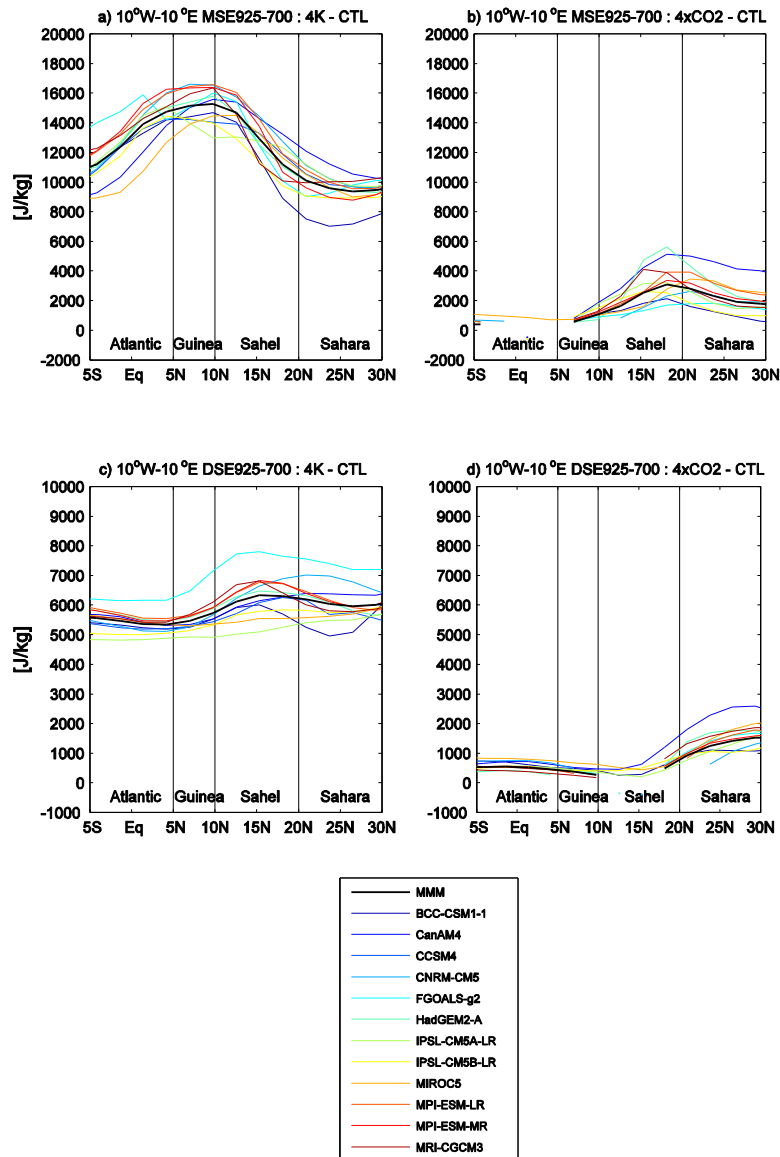
845

846 Figure 9: JAS velocity potential [ $\text{m}^2/\text{s}$ ] (contours and shadings) and divergent wind (vectors)  
 847 at 200 hPa: (a) MMM of the CTL simulation and MMM differences between (b) 4K and (c)  
 848 4xCO<sub>2</sub> experiments and CTL simulation. Divergent wind is computed as the gradient of the  
 849 velocity potential. Vectors and shadings indicate differences significant at 95% confidence  
 850 level. Note that reference scales and vectors are not the same for the CTL simulation and the  
 851 sensitivity experiment responses.



852

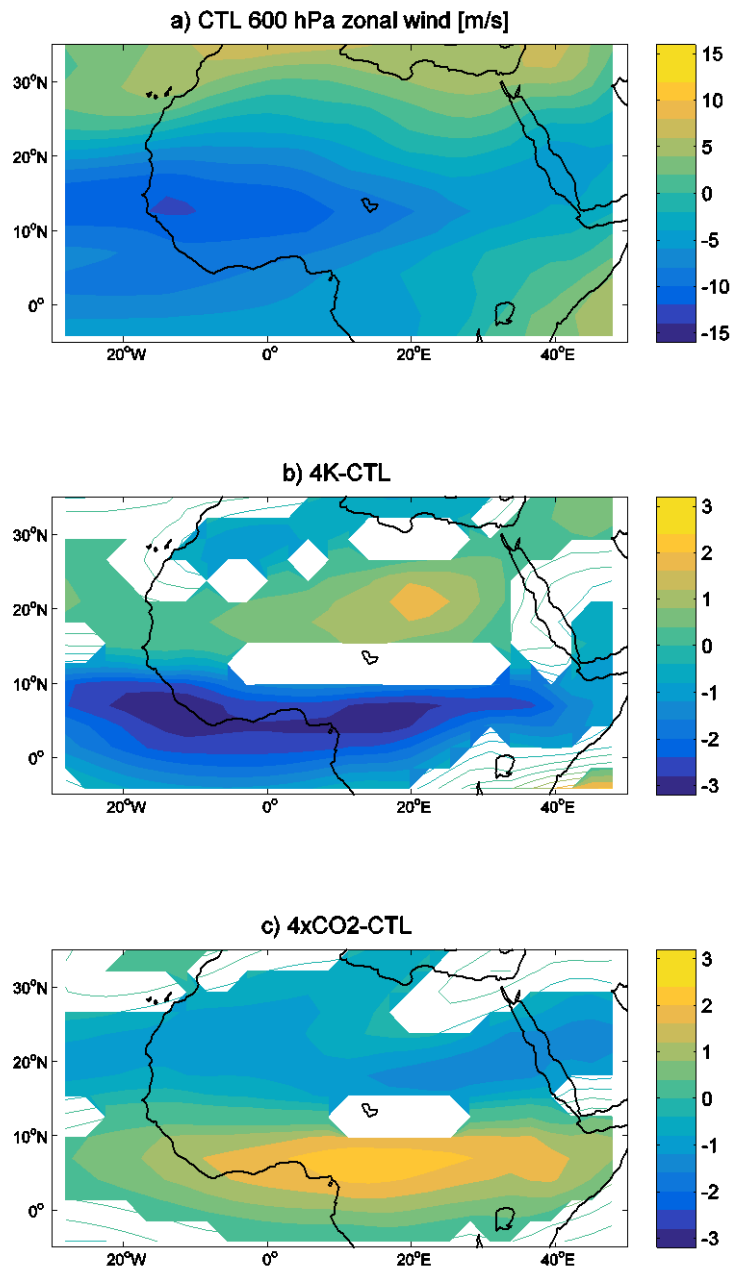
853 Figure 10: JAS 925-700 hPa thickness [m], defined as the difference in geopotential height  
 854 between 700 and 925 hPa: (a) MMM of the CTL simulation and MMM differences between  
 855 (b) 4K and (c) 4xCO2 experiments and CTL simulation. Significant differences at 95%  
 856 confidence level are shaded. Note that the reference scale is not the same for the CTL  
 857 simulation and the sensitivity experiment responses.



858

859 Figure 11: JAS meridional profile of (top) MSE and (bottom) DSE, integrated in the 925-700  
 860 hPa layer [J/kg] and averaged over West Africa [10°W-10°E]: (left) 4K and (right) 4xCO2  
 861 differences with CTL simulation. Significant differences at 95% confidence level are  
 862 displayed. Note the differences in the y-axis values.

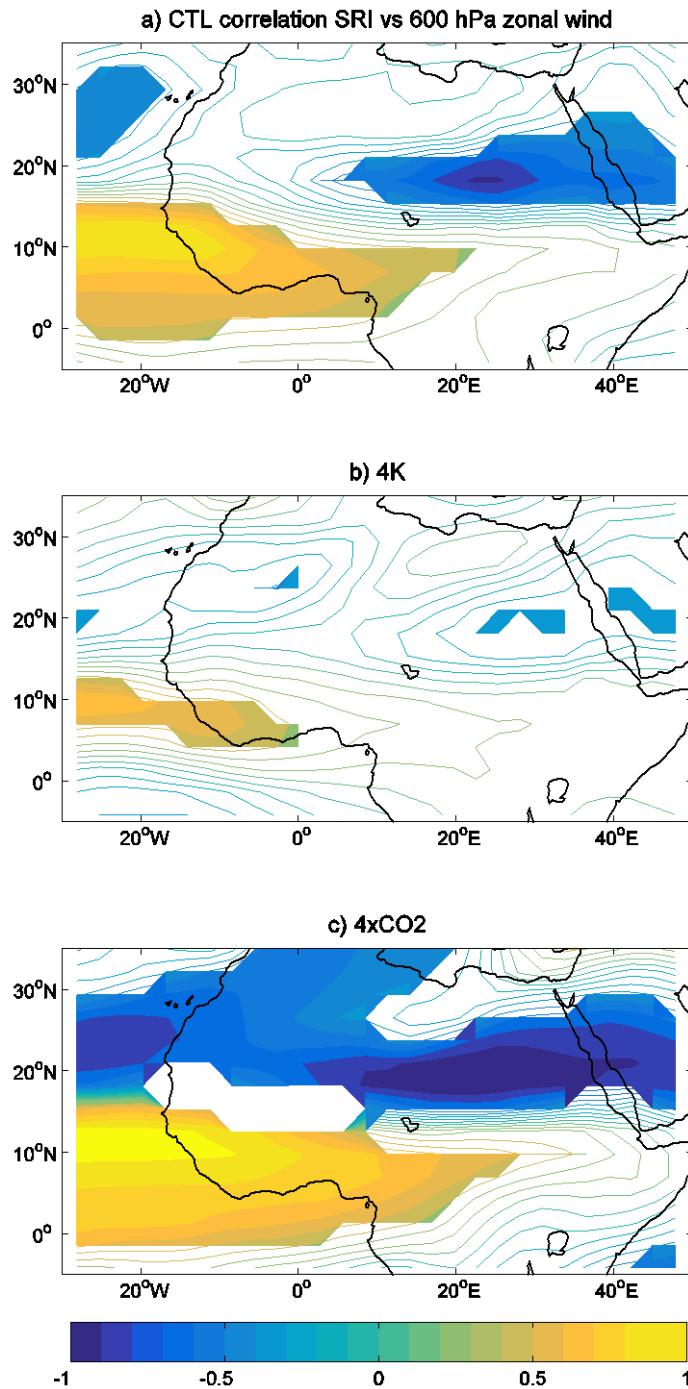
863



864

865 Figure 12: JAS zonal wind at 600 hPa [m]: (a) MMM of the CTL simulation and MMM  
 866 differences between (b) 4K and (c) 4xCO<sub>2</sub> experiments and CTL simulation. Significant  
 867 differences at 95% confidence level are shaded. Note that the reference scale is not the same  
 868 for the CTL simulation and the sensitivity experiment responses.

869

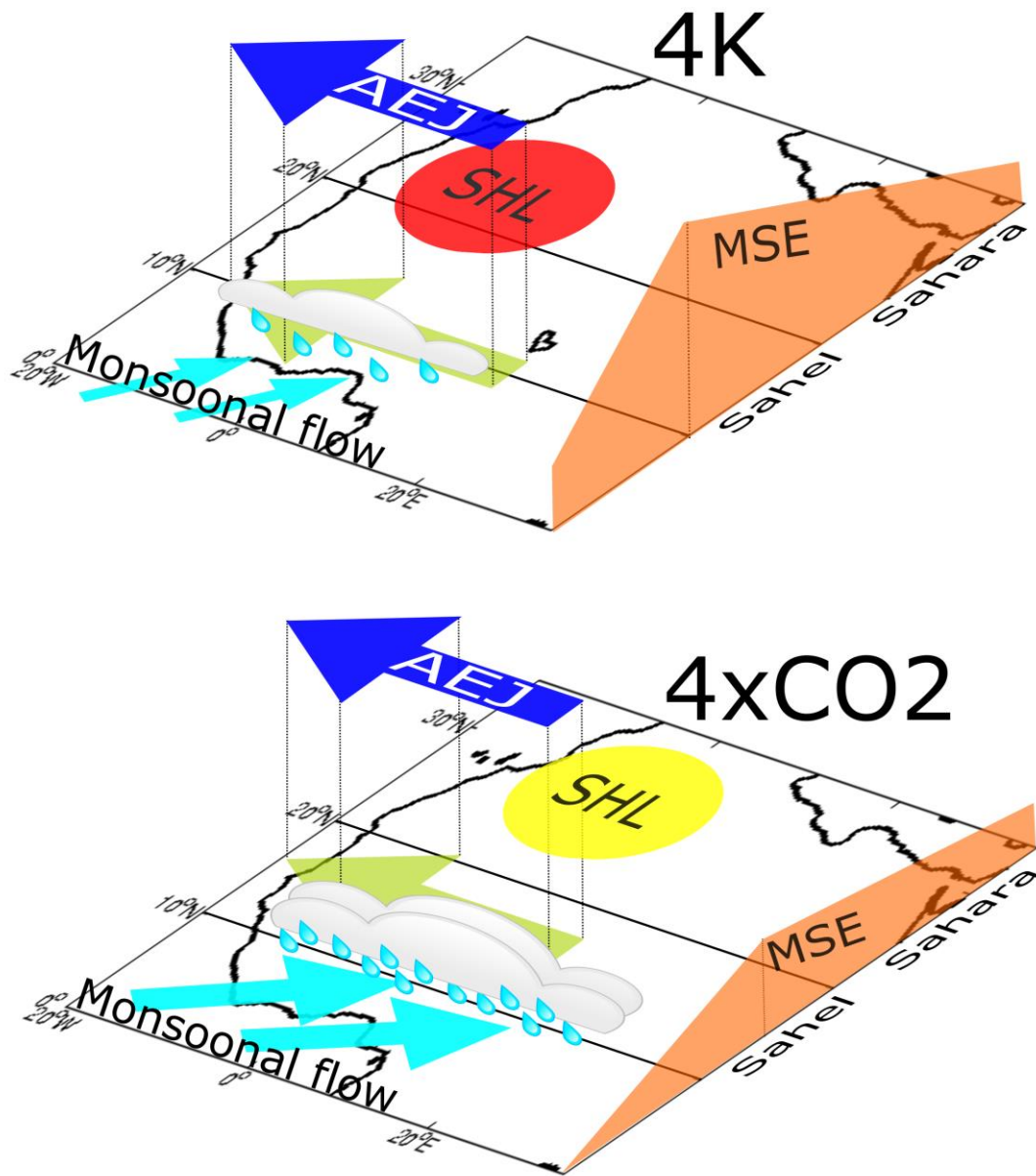


870

871 Figure 13: Correlation between the JAS MMM of SRI and zonal wind at 600 hPa for the  
 872 period 1979-2008: a) CTL simulation, b) 4K and c) 4xCO<sub>2</sub> experiments. Correlations  
 873 significant at 95% confidence level are shaded. The linear trend has been removed from the  
 874 SRI time series and the zonal wind at each grid-point.

875





876

877 Figure 14: Schematics of the WAM dynamics and precipitation in the idealized simulations.

878 The displacement of the monsoonal system in the meridional direction is illustrated, along

879 with the changes in MSE content, monsoonal flow strength, and precipitation amount. A

880 stronger SHL in the 4K experiment is highlighted in red.