

West African monsoon dynamics and precipitation: the competition between global SST warming and CO2 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 CO2 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 Supérieur, 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: <u>marco.gaetani@latmos.ipsl.fr</u>
- 20

21 Abstract

22

Climate variability associated with the West African monsoon (WAM) has important 23 environmental and socio-economic impacts in the region. However, state-of-the-art climate 24 25 models still struggle in producing reliable climate predictions. An important cause of this low predictive skill is the sensitivity of climate models to different forcings. In this study, the 26 27 mechanisms linking the WAM dynamics to the CO2 forcing are investigated, by comparing the effect of the CO2 direct radiative effect with its indirect effect mediated by the global sea 28 29 surface warming. The July-to-September WAM variability is studied in climate simulations extracted from the Coupled Model Intercomparison Project Phase 5 (CMIP5) archive, driven 30 by prescribed sea surface temperature (SST). The individual roles of global SST warming and 31 CO2 atmospheric concentration increase are investigated through idealized experiments 32 simulating a 4 K warmer SST and a quadrupled CO2 concentration, respectively. Results 33 show opposite and competing responses in the WAM dynamics and precipitation. A dry 34 response (-0.6 mm/day) to the SST warming is simulated in the Sahel, with dryer conditions 35 over western Sahel (-0.8 mm/day). Conversely, the CO2 increase produces wet conditions 36 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 that the SST warming affects the Sahelian precipitation through modifications in the global 39 40 tropical atmospheric dynamics, reducing the importance of the regional drivers, while the CO2 increase reinforces the coupling between precipitation and regional dynamics. A general 41 42 agreement in model responses demonstrates the robustness of the identified mechanisms linking the WAM dynamics to the CO2 direct and indirect forcing, and indicates that these 43 44 primary mechanisms are captured by climate models. Results also suggest that the spread in future projections may be caused by unbalanced model responses to the CO2 direct and 45 46 indirect forcing.

- 48 **1. Introduction**
- 49

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

59

WAM precipitation variability at time scales from interannual to multidecadal is mainly 60 driven by global ocean SST anomalies [Giannini et al., 2003]. At the interannual time scale, 61 the Tropical Atlantic SST variability modulates the monsoonal circulation through the 62 meridional land-sea thermal gradient, affecting the meridional displacement of the 63 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 al., 2010; Gaetani et al., 2010]. Moreover, SST variability in the tropical Pacific may induce 66 67 stationary Kelvin and Rossby waves propagating along the Equator and interacting over the Sahel [Rowell, 2001]. The interannual regional and remote SST connections are not 68 69 stationary and are modulated by decadal and multidecadal SST anomalies [Fontaine et al., 2011]. The global ocean signals which dominate the low frequency precipitation variability in 70 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 Mohino, 2015]. On the other hand, positive phases of AMV, by displacing northward the 75 intertropical convergence zone (ITCZ), favour precipitation in the Sahel [Zhang and 76 Delworth, 2006; Ting et al., 2009; Mohino et al., 2011]. A broad and up-to-date review of the 77 78 literature on the SST influence on the climate variability in West Africa is reported by [Rodríguez-Fonseca et al., 2015]. 79

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

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

a response to the weaker net radiative cooling of the atmosphere associated with higher 115 atmospheric CO2 levels affecting the strength of the atmospheric vertical motions. Therefore, 116 the competition between the response of the land-atmosphere system to the local GHG 117 radiative forcing, and the response mediated through the warming of the global SST 118 [Giannini, 2010; Skinner et al., 2012], emerges as a fundamental component of the WAM 119 dynamics and variability, and different sensitivities of climate models to each process appear 120 as a possible explanation of the spread in the projections. Moreover, *Biasutti et al.* [2009] 121 highlighted that the relationship between the regional atmospheric circulation, particularly the 122 123 thermal low over the Sahara desert, and Sahel precipitation is generally well reproduced by the coupled models, both in the 20th and the 21st century, while the same models show 124 discrepancies in future projections. The authors argue that the spread in predictions could be 125 also explained through different abilities in reproducing the mechanisms which influence the 126 regional atmospheric circulation. 127

128

In this context, the open issues regarding the WAM dynamics and precipitation variability 129 can be conveyed in two main arguments. The first one concerns the detailed comprehension 130 of the physical mechanisms connecting the direct and indirect GHG forcing to the WAM 131 132 dynamics, and the specific roles of the large scale and regional atmospheric dynamics in modulating this linkage. The second one concerns the assessment of the ability of climate 133 134 models to simulate the identified physical mechanisms. The objective of this study is to address these issues, by investigating the specific influence of the CO2 atmospheric 135 136 concentration on the WAM dynamics. The direct effect of CO2 concentration increase, and its indirect effect mediated by the global SST warming, are analysed, with a specific focus on 137 138 the atmospheric dynamics response at the regional and global scale. To this aim, multi-model SST-driven sensitivity experiments are analysed, simulating the present climate under 139 140 idealized warmer Global Ocean and higher CO2 atmospheric concentration, respectively. To allow a clear separation of the climate system response, and assess the model sensitivity to 141 the diverse forcings, extreme conditions are imposed in the sensitivity experiments. SST-142 driven simulations are preferred to coupled simulations to minimize the problems affecting 143 the coupled models in correctly locating the Sahelian monsoonal precipitation, due to SST 144 biases in the equatorial Atlantic [Roehrig et al., 2013]. 145 146

147 The paper is structured as follow: models and data analysed are presented in Section 2; the148 response of monsoonal precipitation and circulation in the idealized experiments is presented

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

151

152 **2. Model simulations**

153

The JAS WAM variability and dynamics are studied by analysing numerical experiments 154 from a set of 12 state-of-the-art atmospheric-ocean coupled models selected in the CMIP5 155 archive [Taylor et al., 2012] (see Table 1 for details on the models). In the control simulation 156 157 (CTL), the models are run in atmospheric-only configuration, with prescribed observed SST and sea ice for the period 1979 to the present. Moreover, the simulation includes the observed 158 evolution in the atmospheric composition (including CO2), due to both anthropogenic and 159 natural influences; and the changes in solar forcing, emissions or concentrations of aerosols, 160 and land use. The CTL simulation is the baseline for two sensitivity experiments run either by 161 prescribing uniform 4 K increase in global SST (4K experiment), or by quadrupling the CO2 162 atmospheric concentration while maintaining the SST unchanged. (4xCO2 experiment). The 163 experimental setup is described in detail in Taylor et al. [2012]. Imposing extreme idealized 164 forcing in the sensitivity experiments allows to maximise the impacts and extract the slow 165 166 response of the climate system to the global SST warming (in 4K), and the fast response to the local direct CO2 radiative forcing (in 4xCO2), respectively. The 4xCO2 and 4K idealized 167 168 forcings can be considered extreme in comparison with the observed changes in CO2 concentration (+40% since 1750) and SST (+0.7 K in the period 1880-2012) from the pre-169 170 industrial through the historical period [IPCC, 2014]. However, the idealized scenarios used in this study are comparable with the situation expected in 2100 in the RCP8.5 emission 171 172 scenario, with the CO2 concentration augmented from 390 ppm in 2011 to more than 1000 ppm (more than +260%), and more than 3 K global SST warming [IPCC, 2014]. RCP8.5 is 173 an extreme scenario with a radiative forcing reaching 8.5 W/m⁻² in 2100 [*Riahi et al.*, 2011]. 174 175

To ensure that multi-model mean results are not biased towards models for which more
realisations are available, for each model only one realisation of each experiment is used.
This "one model-one vote" concept [*Santer et al.*, 2009] has been proved successful for
detection of Sahel rainfall significant anomalies in CMIP3 and CMIP5 climate scenarios
[*Monerie et al.*, 2012]. The selection of models is based on the experiments availability,
considering only models for which the three experiments are available. The differences in
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 averaging precipitation and atmospheric fields in JAS. The response of the WAM 184 precipitation and dynamics to idealized conditions is studied by comparing the sensitivity 185 experiments to the CTL simulation, and significant differences in means are detected through 186 a Student's t-test at 95% confidence level. A comprehensive evaluation of the individual 187 models' performances is beyond the scope of the paper. However, the biases of the models 188 for some variables describing the WAM dynamics are presented and discussed in the 189 Supplementary Material, along with the response of the individual models in the sensitivity 190 191 experiments.

192

3. Monsoonal circulation and precipitation response

194

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

- 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
- study, which is focused on the understanding of the WAM dynamics response in idealized
- 221 climate conditions, rather than on making quantitative predictions.
- 222

Results for 4K experiment (Figure 1d) show a dry response in the Sahel to SST warming, 223 with dryer conditions over western Sahel. On the contrary, 4xCO2 experiment (Figure 1e) 224 225 shows wet conditions in the Sahel, with the strongest response to CO2 increase over centraleastern Sahel. These responses to idealized conditions are in line with previous results on the 226 influence of SST warming [Held et al., 2005; Cook and Vizy, 2006; Caminade and Terray, 227 2010] and GHG increase [Haarsma et al., 2005; Hoerling et al., 2006; Giannini, 2010] on 228 WAM precipitation. The modifications in the precipitation patterns are detailed for the 229 individual models by computing a regional rainfall index for the Sahel, the Sahelian rainfall 230 index (SRI) in the domain [10°-20°N, 20°W-30°E], and two sub-regional indices for western 231 [20°W-0°E] and eastern [0°-30°E] Sahel. In the 4K simulation (Figure 2a), all the models 232 show significant negative SRI differences with CTL, the MMM difference is -0.6 mm/day, 233 234 and the multi-model spread ranges between -1.1 and -0.3 mm/day (0.3 mm/day STD). In western Sahel (Figure 2b), the 4K MMM response decreases to -0.8 mm/day, and differences 235 236 are significant in all the models, comprised in [-1.7, -0.2] mm/day (0.5 mm/day STD). In eastern Sahel (Figure 2c), differences are significant in 10 out of 12 models, ranging in [-1.0, 237 238 0.0] mm/day (0.3 mm/day STD), with a -0.5 mm/day MMM difference. Conversely, the SRI response in the 4xCO2 MMM is +0.5 mm/day, the multi-model spread ranges from +0.2 to 239 240 +0.9 mm/day (0.2 mm/day STD), and the differences are significant in 11 out of 12 models (Figure 2a). In western Sahel (Figure 2b), the 4xCO2 MMM response decreases to +0.3 241 mm/day, and differences are significant in 5 out of 12 models, ranging in [0.0, +0.8] mm/day 242 (0.3 mm/day STD). In eastern Sahel (Figure 2c), the 4xCO2 MMM response increases to 243 +0.7 mm/day, and differences are significant in 11 out of 12 models, ranging in [+0.2, +1.2] 244 mm/day (0.3 mm/day STD). The substantial multi-model agreement on the sign of the 245 246 precipitation responses confirms the robustness of the simulated precipitation patterns in response to the SST and CO2 idealized forcings. However, the spread in the multi-model 247 response suggests possible higher sensitivity in individual models to one of the two forcings. 248 The relative importance of the SST warming and CO2 increase in influencing the Sahelian 249 250 precipitation is assessed by computing the sensitivity of the SRI to the global surface

- 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 4xCO2
- experiment, sensitivity is +1.27 mm/day/K in the MMM. This is, in absolute value, 9 times
- higher than in 4K experiment, showing -0.14 mm/day/K in the MMM. A prominent influence
- of the direct CO2 forcing on the Sahelian precipitation is evident, although differences in the
- individual model sensitivity are sizable, ranging from +0.53 mm/day/K in FGOALS-g2 (3
- times higher than in the 4K experiment, -0.16 mm/day/K) to +2.17 mm/day/K in HadGEM2-
- A (24 times higher than in the 4K experiment, -0.09 mm/day/K).
- 259

In addition, it is interesting to notice how, from the superimposition of the responses to SST 260 warming and CO2 increase, the main features of the projected changes in Sahelian 261 precipitation emerge, namely dry anomalies over western Sahel and wet conditions over 262 central-eastern Sahel [Biasutti, 2013; Monerie et al., 2013; Roehrig et al., 2013]. A 263 descriptive illustration is presented in Figure 3, where the 4K-CTL and 4xCO2-CTL 264 precipitation differences (Figures 1d, e) are combined and compared with the 21st century 265 projection simulated by the same models, run in the coupled setup, for the RCP8.5 emission 266 scenario. The combination of the 4K and 4xCO2 responses (Figure 3a, computed as the sum 267 268 of the precipitation patterns displayed in Figures 1d, e) exhibits positive precipitation anomalies in central Sahel (up to around +0.5 mm/day) and the Tropical Atlantic around 5°N 269 270 (up to around +1 mm/day), and dry anomalies in western Sahel (down to around -2 mm/day) and the eastern part of the Guinean coast (down to around -0.5 mm/day). In the RCP8.5 271 272 scenario, at the end of the 21st century, wet anomalies are expected in central-eastern Sahel (up to around +1 mm/day) and the Tropical Atlantic around 5°N (up to around +2 mm/day), 273 274 and dry anomalies are projected in western Sahel (down to around -0.5 mm/day) and the Guinea coast (down to around -1.5 mm/day). Whereas the anomaly patterns are different in 275 276 amplitude and local features, the main spatial structures appear similar. This suggests that the future projected pattern may result from the competition between the different actions of the 277 CO2 forcing on the land-atmosphere system. On the one hand, the indirect action conveyed 278 by the SST warming, which heats the troposphere, imposing stability and reducing deep 279 280 convection over land [Caminade and Terray, 2010]. On the other hand, the direct action on the local radiative balance, producing a weaker net radiative cooling of the atmosphere, a 281 warming of the land surface and a change in the land-sea thermal gradient, which is 282 favourable to the strengthening of large-scale upward motions [Bony et al., 2013]. The 283 evidence of the contrasting effects of the CO2 forcing on the WAM precipitation, along with 284

the different sensitivity shown by individual models, point out the importance of

understanding the balance between the competing actions of the SST warming and CO2

increase and improving the climate model ability in correctly simulate this balance. It is also

- worth noticing that the superimposition pattern in Figure 3 is consistent with the precipitation
- anomalies expected during positive phases of the SHL [*Lavaysse et al.*, 2010b; *Evan et al.*,
- 290 2015].
- 291

The monsoonal flow in the lower troposphere is depicted in Figure 4 through the horizontal 292 293 wind field and the divergence of the moisture transport at 925 hPa. The CTL simulation shows the intertropical discontinuity (ITD) between the southwesterly moist flow from the 294 tropical Atlantic and the northerly dryer flow crossing the Sahara [Issa Lélé and Lamb, 2010], 295 and the associated moisture convergence over the Sahel (Figure 4a). The response of the 296 models to the SST warming is a general weakening of the monsoonal circulation, with no 297 significant changes over the Sahel, accompanied by a slight southward displacement of the 298 moisture transport convergence (Figure 4b). The shift of the latitudinal position of the 299 moisture transport convergence over West Africa is assessed by identifying the mean position 300 301 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 spread ranges between -1.8° and $+1.0^{\circ}$ (0.7° STD). Given the small modifications in the wind 303 304 field, the changes in the moisture convergence can be attributed to the general moistening of the lower troposphere produced by the SST warming (not shown). On the contrary, the 305 306 massive injection of CO2 in the atmosphere produces a sizable strengthening of the westerly flow from the Atlantic across the Sahel, and a slight reinforcement of the northerly flow from 307 308 the Mediterranean, resulting in a clear northward displacement of the moisture convergence belt over the Sahel (Figure 4c). The shift is +1.1° in the MMM, is positive in 11 out of 12 309 310 models, ranging in $[0.0^{\circ}, +3.6^{\circ}]$ (0.9° STD).

311

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

313

314 The competition between the direct and indirect effects of CO2 concentration increase in

modifying circulation and precipitation in the global Tropics has been investigated by *Bony et*

al. [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

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 Figure 5, the global net radiative balance, defined at the top of the atmosphere as the 320 difference between the downward shortwave radiation and the sum of upward shortwave and 321 longwave radiation, is presented for CTL and sensitivity experiments. The CTL experiment 322 (Figure 5a) represents the warming of the atmosphere in the Tropics and sub-Tropics, with 323 cooling over the desert belt in the Northern Hemisphere, and in the extra-Tropics (north of 324 45-50°N and south of 10-15°S). In the 4K simulation (Figure 5b), a generalized net radiative 325 cooling of the atmosphere is found, reaching -16 W/m^2 in the Sahara, which is ascribed to the 326 increase in losses of longwave radiation associated to the surface warming (not shown). On 327 the contrary, the increased CO2 concentration produces a net radiative warming in 4xCO2 328 (Figure 5c), reaching $+14 \text{ W/m}^2$ in the Sahara, which is associated with reduced losses in 329 both longwave and shortwave radiation (not shown). 330

331

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

335

$MSE = gz + C_pT + Lq,$

336 where gz is the geopotential energy (with g the gravitational acceleration and z the geopotential height), C_pT the enthalpy (with C_p the specific heat of dry air at constant 337 338 pressure and T the temperature), and Lq is the latent energy associated with evaporation and condensation of water (with L the latent heat of evaporation and q the specific humidity). As 339 340 transformation of enthalpy and latent energy available in the lower troposphere into geopotential energy in the upper levels is the main signal of convection, MSE is directly 341 342 related to monsoonal precipitation [Fontaine and Philippon, 2000]. The DSE accounts for only enthalpy and geopotential energy. The differences between sensitivity and CTL 343 experiments are presented in Figure 6. The CTL simulation shows a maximum in the tropical 344 belt, which is associated to the latent heat content, while DSE is more important over the 345 subtropical continental surfaces (Figure 6a, b). The surface warming in 4K produces a global 346 and almost uniform increase of MSE, with a maximum over the northern Tropics, which is 347 348 associated to the changes in the latent heat content (Figure 6c, d). On the contrary, the response to increased CO2 concentration in 4xCO2, although weaker, shows a clear 349 350 relationship with the changes in the DSE content, resulting in a meridional gradient and well localized maxima over the Sahel, the Arabian Peninsula, southwestern and eastern Asia, and 351 North America (Figure 6e, f). It appears that the response of MSE to the SST warming is 352

mostly associated to the change of moist component at the global scale, while the response in the 4xCO2 experiment is driven by the modifications in the dry component localized over the northern continental masses.

356

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

363

The CTL simulation captures the main features of the meridional circulation over southern 364 Atlantic and West Africa, illustrating the northern and southern overturning cells associated 365 to the meridional Hadley circulation, along with the deep convection associated to the ITCZ 366 around 10°N and the convergence in the lower troposphere over the Sahel associated to the 367 monsoonal circulation (Figure 7a). By comparing CTL with the sensitivity experiments, it is 368 369 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 4xCO2 the southern overturning cell is reinforced along with deep convection over the Sahel, and the 371 372 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 with the WAM dynamics and induced by the oceanic warm pools in the Indian Ocean and in 375 376 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 summer monsoon (Figure 8a). The TEJ intensity is strongly related to the precipitation 379 variability, wet conditions in the Sahel being concomitant with stronger than average TEJ and 380 vice versa [Nicholson, 2013]. The sensitivity experiments show anomalous ascending 381 382 motions as a response to SST warming over the western Pacific and anomalous subsidence over Africa and Indian Ocean, along with a weakening of the TEJ (Figure 8b). The change in 383 the TEJ intensity is assessed over West Africa by averaging the zonal wind at 200 hPa in the 384 domain [5°-15°N, 20°W-30°E]. It results that, in the 4K MMM, the TEJ is 4.7 m/s weaker 385 than in the CTL run, and the multi-model response ranges in [+0.5, +10.1] m/s (2.8 m/s 386

STD). Conversely, the 4xCO2 experiment simulates weaker upward motions in the western
and eastern Pacific, and an anomalous organized zonal cell over the Africa-Atlantic sector,
with a westerly flux in the lower troposphere into West Africa (see also Figure 4c) and a
stronger TEJ (Figure 8c). In the 4xCO2 experiment, the MMM response is -2.9 m/s, and the
multi-model response ranges in [-3.75, -0.12] m/s (0.97 m/s STD).

392

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

400 the teleconnection over Africa in 4xCO2 (Figure 9c).

401

The analysis of the response of the global dynamics fields and the modifications in the Sahelian precipitation connections suggests that the increase in CO2 concentration favours the meridional circulation and the zonal connectivity in the tropical belt, by forcing a meridional energy gradient. Meridional circulation and zonal connectivity are conversely weakened by a warmer Global Ocean, which produces an energy excess in the tropical belt.

407

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

409

In this section the impact of the SST and CO2 idealized forcings on the WAM regional
drivers is discussed, by analysing the modifications in the most relevant features of the
atmospheric circulation over West Africa, and the changes in their relationship with the
Sahelian precipitation.

414

The response of the SHL is first analysed, by computing the 925-700 hPa thickness (Z925-700), which measures the thermal expansion of the lower troposphere [*Lavaysse et al.*, 2009] The CTL simulation well reproduces the climatological location of SHL over western Sahara (Figure 10a). The 4K experiment is characterized by a general thickening of the lower troposphere compared to CTL (Figure 10b), which is consistent with the global warming of the surface associated with the forcing SST pattern (not shown). A peak in thickness is 421 evident along 15°N (around +60 m), which inhibits the penetration of the monsoonal flow and rainfall inland compared to CTL (see Figure 1d). The SHL latitudinal position is 422 identified as the mean position of the Z925-700 maximum in the longitudinal domain [10°W, 423 10°E]. In the 4K simulation, 10 out of 12 models show a southward shift, and the multi-424 model spread ranges in $[-2.0^{\circ}, +0.6^{\circ}]$, $(0.7^{\circ} \text{ STD})$, resulting in a -0.7° meridional 425 displacement in the MMM. On the contrary, the 4xCO2 experiment presents a significant 426 warming over land in the boreal summer hemisphere, resulting in thermal anomalies over the 427 continental surface in North America, Eurasia and the Sahara (not shown). As a consequence 428 429 of the Sahara desert warming, thickness over West Africa shows a meridional gradient (from around 0 m in the Sahel to around +17 m along the Mediterranean southern coast, see Figure 430 10c), which is favourable to the northward displacement of the precipitation belt (see Figure 431 1e). The SHL displacement is $+1.0^{\circ}$ northward in the 4xCO2 MMM, comprised in [0.0°, 432 $+2.6^{\circ}$ (0.9° STD) in the multi-model ensemble. The comparison between 4K and 4xCO2 433 changes in thickness shows a larger thickening in 4K everywhere in West Africa, which is 434 associated with reduced precipitation, pointing to the importance of the regional pattern of the 435 modifications, in spite of the mere amplitude. 436

437

438 The modifications in the MSE and DSE content in the lower troposphere are detailed for West Africa by computing the meridional profile in the domain $[10^{\circ}W-10^{\circ}E]$, to emphasize 439 440 the land-sea contrast between West Africa and southern Atlantic (Figure 11). Differences between sensitivity experiments and CTL show a general and significant MSE increase in all 441 442 the 4K models, from tropical Atlantic to Sahara and peaking over the coast of Guinea (Figure 11a). On the contrary, the modifications simulated in the 4xCO2 experiment are in general 443 444 smaller, with all the models simulating significant and positive differences over the African continent and peaking over the Sahel, and no significant differences over the ocean (Figure 445 11b). When DSE is selected (Figure 11c, d), it results that the meridional profile is dominated 446 in both the experiments by the latent heat associated with evaporation. Enthalpy over Sahara 447 desert is important in maintaining the MSE meridional gradient in 4xCO2, but in 4K it is not 448 enough to produce a significant meridional gradient. In spite of the larger availability of MSE 449 in the 4K experiment compared to 4xCO2, the simulated profile is less favourable to the 450 inland progression of the monsoonal circulation, because of the peak around the Guinea 451 coast, which prevents the monsoonal flow to penetrate inland (Figure 11c). In comparison 452 with the CTL simulations, the maximum of the MSE meridional profile is shifted 1.2° 453 southward in the MMM, comprised between -2.4° and 0.0° (0.7° STD) in the multi-model 454

455 ensemble. On the contrary, the 4xCO2 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 $[+0.0^{\circ}, +1.6^{\circ}]$ (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

The zonal wind at 600 hPa is analysed to portray the African easterly jet (AEJ), which is a 461 key element in the WAM dynamics. The AEJ develops in response to the meridional thermal 462 463 gradient over West Africa, with an east-west orientation along 12°N [Cook, 1999], and is maintained by the meridional circulations associated with the monsoon, namely the Hadley 464 circulation and the SHL [Thorncroft and Blackburn, 1999; Chen, 2005]. The AEJ variability 465 influences the intensity and position of the WAM precipitation pattern, with a stronger 466 (weaker) than average jet associated to negative (positive) rainfall anomalies over Sahel, and 467 a northern (southern) displacement associated to a more (less) inland penetrating monsoonal 468 flow [Cook, 1999]. Moreover, AEJ may modulate the WAM precipitation through the 469 synoptic activity of the African easterly waves (AEW) on its southern flank, along 10°N 470 [Diedhiou et al., 1999; Fink, 2003]. The main axis of the AEJ is correctly reproduced in the 471 472 CTL simulation (Figure 12a). The 4K experiment shows stronger (weaker) easterly winds south (north) of 10°N, corresponding to a southern displacement of the AEJ, while the 473 474 opposite pattern is found in the 4xCO2 experiment (Figure 12b, c). The modifications in the AEJ location in the sensitivity experiments are estimated by determining the mean latitudinal 475 476 position of the minimum of the zonal wind at 600 hPa in the longitudinal domain [20°W, 30°E]. As a result, the AEJ shifts southward by -3.0° in the 4K MMM, and the multi-model 477 478 spread ranges is $[-5.3^{\circ}, -0.2^{\circ}]$ (1.5° STD). In the 4xCO2 experiment, the shift in the MMM is $+1.2^{\circ}$, ranging in $[+0.0^{\circ}, +2.5^{\circ}]$ (0.8° STD). The modifications in the AEJ dynamics are 479 480 consistent with the precipitation changes displayed in Figure 1 and also a linkage with the meridional energy profiles in Figure 11 appears evident, being the stronger enthalpy gradient 481 more favourable to a northward displaced jet in the 4xCO2 experiment. 482

483

484 The linkage between the Sahelian precipitation and the thermal forcing over West Africa is

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 4xCO2 (significant in 9 out of 12, and 10 out of 12 models, respectively), with 4xCO2 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

The coherence between precipitation and regional monsoonal circulation is studied by 494 495 correlating the SRI and the West African Monsoon Index (WAMI), which is defined in the domain [20°W-20°E, 3°-13°N] as the difference between the standardized wind modulus at 496 497 850 hPa and the standardized zonal wind at 200 hPa [Garric et al., 2002]. The WAMI is used as an indicator of the intensity of the regional monsoonal cell, and is positively correlated to 498 the Sahel precipitation at the interannual time scale [Fontaine et al., 1995]. GPCP data and 499 ERAI reanalysis confirm the correlation for the period 1979-2008 (r = 0.77), and CTL 500 simulations partially reproduce the correlation (r = 0.46 in the MMM), with 7 out of 12 501 models showing significant positive correlations (Table 2). Interestingly, in the 4K 502 experiment the relationship is even less robust than in CTL (r = 0.34 in the MMM, significant 503 504 in 4 out of 12 models), while robustness is recovered in 4xCO2 (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 Sahelian precipitation are studied by analysing the correlation between SRI and AEJ, which 509 510 is represented through the zonal wind at 600 hPa. The positive correlation between precipitation and intensity and meridional displacement over West Africa of the AEJ is 511 512 illustrated in the CTL simulation, showing positive (negative) correlations along the 10°N (18°N) axis (Figure 13a). When the correlation is computed in the sensitivity experiments, the 513 general pattern is conserved in both 4K and 4xCO2, but significance is strongly reduced in 514 the former, while the signal is robust in the latter, where the axis over West Africa is also 515 better represented (Figure 13b, c). 516

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 CO2 concentration increase is particularly

521 effective in modifying the dynamics over the Sahara desert, favouring the mechanisms

sociated with wet conditions in the Sahel. Furthermore, while the SST warming appears to

weaken the linkage between precipitation and regional dynamics, the CO2 increase clearlystrengthens the coupling.

525

526 6. Summary and conclusions

527

In this study, the mechanisms linking the WAM dynamics to the CO2 forcing are 528 investigated. To this aim, extreme idealized conditions are imposed in multi-model climate 529 simulations to maximise and clearly separate the response of the WAM dynamics to the 530 531 opposite and competing effects induced by CO2. On the one hand, the effect of the direct radiative forcing, which tends to reinforce the monsoonal circulation, is studied in sensitivity 532 experiments run with prescribed observed SST and quadrupled CO2 concentration. On the 533 other hand, the indirect CO2 effect mediated by the global SST warming, which weakens the 534 monsoon, is studied in sensitivity experiments where the global SST is uniformly increased 535 536 by 4 K.

537

The precipitation response over Sahel shows opposite modifications, with dry conditions associated with the SST warming (-0.6 mm/day in the Sahel, down to -0.8 mm/day in western Sahel), while increased CO2 concentration leads to wet anomalies (+0.5 mm/day in the Sahel, up to +0.7 mm/day in eastern Sahel). Moreover, the modifications in the precipitation patterns appear associated with coherent modifications in the monsoonal circulation, namely in the meridional thermal forcing, the monsoonal flow in the lower troposphere, the AEJ and 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 scale, weakening the connection between precipitation and the regional drivers. A warmer 547 Global Ocean reinforces evaporation in the Tropics, resulting in an inhibition of the 548 monsoonal flow and the ITCZ migration over West Africa. Conversely, the CO2 increase 549 strengthens the coupling between precipitation and the regional dynamics over West Africa. 550 The local atmospheric radiative forcing, resulting from the direct action of the CO2 increase, 551 552 warms the Sahara and enhances evaporation over the Sahel, producing specific humidity anomalies in the low troposphere. The consequent increase in the low tropospheric latent heat 553 554 and enthalpy content over Sahara and Sahel reinforces the MSE meridional gradient over West Africa, leading to more intense monsoonal circulation and precipitation. More 555 specifically, the westerly flow over the Sahel is reinforced, accompanied by a northward shift 556

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

565

The general agreement among models in their response to the idealized conditions 566 demonstrates the robustness of the identified mechanisms linking the WAM dynamics to the 567 SST and CO2 idealized forcings, and reflects the fact that the simulation of these mechanisms 568 does not depend on the model physics, nor on the model performance. This robustness also 569 confirms the value of using idealized forcings to maximise and separate the response of the 570 climate system. Furthermore, the similarity of the future projections in the RCP8.5 scenario 571 with the combined responses in the 4K and 4xCO2 simulations, though only qualitative, 572 573 clearly shows how the competition between the global SST warming and the CO2 increase is 574 a key factor in driving the WAM variability and change. Indeed, it is argued that the large spread in the simulation of the WAM dynamics and precipitation may be related to an 575 576 unbalanced model response to the direct and indirect CO2 forcings, i.e., dry (wet) biases may be produced by specific model sensitivity to the SST warming (CO2 increase). The analysed 577 578 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 CO2 forcings on the WAM dynamics, and the assessment of the model ability to reproduce the amplitude of this impact, still need to be addressed. This could be achieved by 581 analysing model simulations of periods well covered with observations and one of the two 582 mechanisms is clearly dominant. In this respect, the 20th century and/or the mid-Holocene 583 could be optimal candidates to analyse the processes responding to changes in the land-sea 584 contrast, and this will be the subject of a forthcoming study. 585

586

587 Acknowledgements

588

This work benefitted from the support of the Agence Nationale de la Recherche (ANR) grant
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
- 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
- useful discussions, and two anonymous reviewers for their insightful comments that have
- 599 greatly improved the quality of the paper.
- 600

601 **References**

- Adler, R. F. et al. (2003), The Version-2 Global Precipitation Climatology Project (GPCP)
 Monthly Precipitation Analysis (1979–Present), *J. Hydrometeorol.*, *4*, 1147–1167,
 doi:10.1175/1525-7541(2003)004<1147:TVGPCP>2.0.CO;2.
- Bader, J., and M. Latif (2003), The impact of decadal-scale Indian Ocean sea surface
 temperature anomalies on Sahelian rainfall and the North Atlantic Oscillation, *Geophys. Res. Lett.*, 30, 1–4, doi:10.1029/2003GL018426.
- Biasutti, M. (2013), Forced Sahel rainfall trends in the CMIP5 archive, J. Geophys. Res.
 Atmos., 118, 1613–1623, doi:10.1002/jgrd.50206.
- Biasutti, M., A. H. Sobel, and S. J. Camargo (2009), The Role of the Sahara Low in
 Summertime Sahel Rainfall Variability and Change in the CMIP3 Models, *J. Clim.*, 22,
 5755–5771, doi:10.1175/2009JCLI2969.1.
- Bony, S., G. Bellon, D. Klocke, S. Sherwood, S. Fermepin, and S. Denvil (2013), Robust
 direct effect of carbon dioxide on tropical circulation and regional precipitation, *Nat. Geosci.*, 6, 447–451, doi:10.1038/ngeo1799.
- Caminade, C., and L. Terray (2010), Twentieth century sahel rainfall variability as simulated
 by the ARPEGE AGCM, and future changes, *Clim. Dyn.*, *35*, 75–94,
 doi:10.1007/s00382-009-0545-4.
- Chauvin, F., R. Roehrig, and J. P. Lafore (2010), Intraseasonal variability of the saharan heat
 low and its link with midlatitudes, *J. Clim.*, 23, 2544–2561,
 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.
- Cook, K. H. (1999), Generation of the African easterly jet and its role in determining West
 African precipitation, *J. Clim.*, *12*, 1165–1184, doi:10.1175/15200442(1999)012<1165:GOTAEJ>2.0.CO;2.
- Cook, K. H., and E. K. Vizy (2006), Coupled model simulations of the West African
 monsoon system: Twentieth- and twenty-first-century simulations, *J. Clim.*, *19*, 3681–
 3703, doi:10.1175/JCLI3814.1.
- 630 Diedhiou, a., S. Janicot, a. Viltard, P. De Felice, and H. Laurent (1999), Easterly wave

- regimes and associated convection over West Africa and tropical Atlantic: Results from
 the NCEP/NCAR and ECMWF reanalyses, *Clim. Dyn.*, *15*, 795–822,
- 633 doi:10.1007/s003820050316.
- Evan, A. T., C. Flamant, C. Lavaysse, C. Kocha, and A. Saci (2015), Water Vapor–Forced
 Greenhouse Warming over the Sahara Desert and the Recent Recovery from the
 Sahelian Drought, *J. Clim.*, 28, 108–123, doi:10.1175/JCLI-D-14-00039.1.
- Fink, A. H. (2003), Spatiotemporal variability of the relation between African Easterly
 Waves and West African Squall Lines in 1998 and 1999, *J. Geophys. Res.*, 108, 1–17,
 doi:10.1029/2002JD002816.
- Fontaine, B., and N. Philippon (2000), Seasonal Evolution of Boundary Layer Heat Content
 in the West African Monsoon From the Ncep / Ncar, *Int. J. Climatol.*, 20, 1777–1790.
- Fontaine, B., S. Janicot, and V. Moron (1995), Rainfall anomaly patterns and wind field
 signals over West Africa in August (1958-1989), *J. Clim.*, *8*, 1503–1510.
- Fontaine, B., J. Garcia-Serrano, P. Roucou, B. Rodriguez-Fonseca, T. Losada, F. Chauvin, S.
 Gervois, S. Sijikumar, P. Ruti, and S. Janicot (2010), Impacts of warm and cold
 situations in the Mediterranean basins on the West African monsoon: Observed
 connection patterns (1979-2006) and climate simulations, *Clim. Dyn.*, *35*, 95–114,
 doi:10.1007/s00382-009-0599-3.
- Fontaine, B., M. Gaetani, A. Ullmann, and P. Roucou (2011), Time evolution of observed
 July-September sea surface temperature-Sahel climate teleconnection with removed
 quasi-global effect (1900-2008), *J. Geophys. Res. Atmos.*, *116*, 1–17,
 doi:10.1029/2010JD014843.
- Gaetani, M., B. Fontaine, P. Roucou, and M. Baldi (2010), Influence of the Mediterranean
 Sea on the West African monsoon: Intraseasonal variability in numerical simulations, *J. Geophys. Res.*, *115*, 1–17, doi:10.1029/2010JD014436.
- Garric, G., H. Douville, and M. Déqué (2002), Prospects for improved seasonal predictions of
 monsoon precipitation over Sahel, *Int. J. Climatol.*, 22, 331–345, doi:10.1002/joc.736.
- Giannini, A. (2010), Mechanisms of Climate Change in the Semiarid African Sahel: The
 Local View, J. Clim., 23, 743–756, doi:10.1175/2009JCLI3123.1.
- Giannini, A., R. Saravanan, and P. Chang (2003), Oceanic forcing of Sahel rainfall on
 interannual to interdecadal time scales, *Science*, *302*, 1027–1030,
 doi:10.1126/science.1089357.
- Haarsma, R. J., F. M. Selten, S. L. Weber, and M. Kliphuis (2005), Sahel rainfall variability
 and response to greenhouse warming, *Geophys. Res. Lett.*, *32*, 1–4,
 doi:10.1029/2005GL023232.
- Held, I. M., T. L. Delworth, J. Lu, K. L. Findell, and T. R. Knutson (2005), Simulation of
 Sahel drought in the 20th and 21st centuries., *Proc. Natl. Acad. Sci. U. S. A.*, *102*,
 17891–17896, doi:10.1073/pnas.0509057102.
- Hoerling, M., J. Hurrell, J. Eischeid, and A. Phillips (2006), Detection and attribution of
 twentieth-century northern and southern African rainfall change, *J. Clim.*, *19*, 3989–
 4008, doi:10.1175/JCLI3842.1.
- IPCC, (2014), Climate Change 2014: Synthesis Report. Contribution of Working Groups I, II
 and III to the Fifth Assessment Report of the Intergovernmental Panel on Climate
 Change [Core Writing Team, R.K. Pachauri and L.A. Meyer (eds.)]. IPCC, Geneva,
 Switzerland, 151 pp.

- Issa Lélé, M., and P. J. Lamb (2010), Variability of the Intertropical Front (ITF) and Rainfall
 over the West African Sudan–Sahel Zone, *J. Clim.*, 23, 3984–4004,
 doi:10.1175/2010JCLI3277.1.
- Kandji, S. T., L. Verchot, and J. Mackensen (2006), *Climate change and variability in the Sahel region : impacts and adaptation strategies in the agricultural sector*. UNEP &
 ICRAF.
- Lavaysse, C., C. Flamant, S. Janicot, D. J. Parker, J. P. Lafore, B. Sultan, and J. Pelon (2009),
 Seasonal evolution of the West African heat low: A climatological perspective, *Clim. Dyn.*, *33*, 313–330, doi:10.1007/s00382-009-0553-4.
- Lavaysse, C., C. Flamant, S. Janicot, and P. Knippertz (2010a), Links between African
 easterly waves, midlatitude circulation and intraseasonal pulsations of the West African
 heat low, *Q. J. R. Meteorol. Soc.*, *136*, 141–158, doi:10.1002/qj.555.
- Lavaysse, C., C. Flamant, and S. Janicot (2010b), Regional-scale convection patterns during
 strong and weak phases of the Saharan heat low, *Atmos. Sci. Lett.*, *11*, 255–264,
 doi:10.1002/asl.284.
- Lavaysse, C., J. P. Chaboureau, and C. Flamant (2011), Dust impact on the west african heat
 low in summertime, *Q. J. R. Meteorol. Soc.*, *137*, 1227–1240, doi:10.1002/qj.844.
- Losada, T., B. Rodríguez-Fonseca, S. Janicot, S. Gervois, F. Chauvin, and P. Ruti (2010), A
 multi-model approach to the Atlantic Equatorial mode: Impact on the West African
 monsoon, *Clim. Dyn.*, *35*, 29–43, doi:10.1007/s00382-009-0625-5.
- Lu, J., and T. L. Delworth (2005), Oceanic forcing of the late 20th century Sahel drought,
 Geophys. Res. Lett., 32, L22706, doi:10.1029/2005GL023316.
- Meehl, G. A., C. Covey, K. E. Taylor, T. Delworth, R. J. Stouffer, M. Latif, B. McAvaney,
 and J. F. B. Mitchell (2007), THE WCRP CMIP3 Multimodel Dataset: A New Era in
 Climate Change Research, *Bull. Am. Meteorol. Soc.*, *88*, 1383–1394,
 doi:10.1175/BAMS-88-9-1383.
- Mohino, E., S. Janicot, and J. Bader (2011), Sahel rainfall and decadal to multi-decadal sea
 surface temperature variability, *Clim. Dyn.*, *37*, 419–440, doi:10.1007/s00382-010-08672.
- Monerie, P. A., B. Fontaine, and P. Roucou (2012), Expected future changes in the African
 monsoon between 2030 and 2070 using some CMIP3 and CMIP5 models under a
 medium-low RCP scenario, *J. Geophys. Res. Atmos.*, *117*, 1–12,
 doi:10.1029/2012JD017510.
- Monerie, P.-A., P. Roucou, and B. Fontaine (2013), Mid-century effects of Climate Change
 on African monsoon dynamics using the A1B emission scenario, *Int. J. Climatol.*, *33*,
 881–896, doi:10.1002/joc.3476.
- Nicholson, S. E. (2013), The West African Sahel: A Review of Recent Studies on the
 Rainfall Regime and Its Interannual Variability, *ISRN Meteorol.*, 2013, 32,
 doi:10.1155/2013/453521.
- Nicholson, S. E. et al. (2003), Validation of TRMM and Other Rainfall Estimates with a
 High-Density Gauge Dataset for West Africa. Part I: Validation of GPCC Rainfall
 Product and Pre-TRMM Satellite and Blended Products, *J. Appl. Meteorol.*, *42*, 1337–
 1354, doi:10.1175/1520-0450(2003)042<1337:VOTAOR>2.0.CO;2.
- Panthou, G., Vischel, T. and Lebel, T. (2014), Recent trends in the regime of extreme rainfall
 in the Central Sahel. Int. J. Climatol., 34: 3998–4006. doi:10.1002/joc.3984.

- Riahi, K., S. Rao, V. Krey, C. Cho, V. Chirkov, G. Fischer, G. Kindermann, N. Nakicenovic,
 and P. Rafaj (2011), RCP 8.5-A scenario of comparatively high greenhouse gas
 emissions, *Clim. Change*, *109*, 33–57, doi:10.1007/s10584-011-0149-y.
- Rodríguez-Fonseca, B. et al. (2015), Variability and Predictability of West African Droughts:
 A Review on the Role of Sea Surface Temperature Anomalies, *J. Clim.*, 28, 4034–4060,
 doi:10.1175/JCLI-D-14-00130.1.
- Roehrig, R., D. Bouniol, F. Guichard, F. D. Hourdin, and J. L. Redelsperger (2013), The
 present and future of the west african monsoon: A process-oriented assessment of
 CMIP5 simulations along the AMMA transect, *J. Clim.*, *26*, 6471–6505,
 doi:10.1175/JCLI-D-12-00505.1.
- Rowell, D. P. (2001), Teleconnections between the tropical Pacific and the Sahel, Q. J. R.
 Meteorol. Soc., 127, 1683–1706, doi:10.1002/qj.49712757512.
- Rowell, D. P. (2013), Simulating SST teleconnections to Africa: What is the state of the art?,
 J. Clim., *26*, 5397–5418, doi:10.1175/JCLI-D-12-00761.1.
- Santer, B. D. et al. (2009), Incorporating model quality information in climate change
 detection and attribution studies., *Proc. Natl. Acad. Sci. U. S. A.*, *106*, 14778–14783,
 doi:10.1073/pnas.0901736106.
- Skinner, C. B., M. Ashfaq, and N. S. Diffenbaugh (2012), Influence of Twenty-First-Century
 Atmospheric and Sea Surface Temperature Forcing on West African Climate, *J. Clim.*,
 25, 527–542, doi:10.1175/2011JCLI4183.1.
- Tanaka, H. L., N. Ishizaki, and A. Kitoh (2004), Trend and interannual variability of Walker,
 monsoon and Hadley circulations defined by velocity potential in the upper troposphere, *Tellus, Ser. A Dyn. Meteorol. Oceanogr.*, *56*, 250–269, doi:10.1111/j.16000870.2004.00049.x.
- Taylor, K. E., R. J. Stouffer, and G. a. Meehl (2012), An overview of CMIP5 and the
 experiment design, *Bull. Am. Meteorol. Soc.*, *93*, 485–498, doi:10.1175/BAMS-D-1100094.1.
- Thorncroft, C. D., and M. Blackburn (1999), Maintenance of the African easterly jet, Q. J. R.
 Meteorol. Soc., 125, 763–786, doi:10.1002/qj.49712555502.
- Ting, M., Y. Kushnir, R. Seager, and C. Li (2009), Forced and Internal Twentieth-Century
 SST Trends in the North Atlantic, *J. Clim.*, 22, 1469–1481,
 doi:10.1175/2008JCLI2561.1.
- Trenberth, K.E., P.D. Jones, P. Ambenje, R. Bojariu, D. Easterling, A. Klein Tank, D. Parker,
 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
- Report of the Intergovernmental Panel on Climate Change [Solomon, S., D. Qin, M.
 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.
- Villamayor, J., and E. Mohino (2015), Robust Sahel drought due to the Interdecadal Pacific
 Oscillation in CMIP5 simulations, *Geophys. Res. Lett.*, 42, 1214–1222,
 doi:10.1002/2014GL062473.
- Vizy, E. K., and K. H. Cook (2009), A mechanism for African monsoon breaks:
 Mediterranean cold air surges, J. Geophys. Res. Atmos., 114, 1–19,
- 765 doi:10.1029/2008JD010654.

- Xue, Y. et al. (2010), Intercomparison and analyses of the climatology of the West African
 Monsoon in the West African Monsoon Modeling and Evaluation project (WAMME)
 first model intercomparison experiment, *Clim. Dyn.*, *35*, 3–27, doi:10.1007/s00382-0100778-2.
- 770 Zhang, R., and T. L. Delworth (2006), Impact of Atlantic multidecadal oscillations on
- India/Sahel rainfall and Atlantic hurricanes, *Geophys. Res. Lett.*, 33, L17712,
 doi:10.1029/2006GL026267.
- 773

774 Tables

775

- Table 1: Models analysed. CMIP5 model information and outputs are available through the
- Earth System Grid Federation archive (<u>http://cmip-pcmdi.llnl.gov/cmip5</u>).

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

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

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



Figure 1: JAS precipitation [mm/day]: (a) GPCP observations, (b) MMM of the CTL

simulation and (c) difference with observations, and MMM differences between (d) 4K and

(e) 4xCO2 experiments and CTL simulation. Significant differences at 95% confidence level

are shaded. Note that the reference scale is not the same for the climatology, the model bias,

and the sensitivity experiment responses.



Figure 2: JAS precipitation [mm/day]: differences between the 4K (blue circles) and 4xCO2
(red circles) experiments and the CTL simulation, in (a) Sahel [10°-20°N, 20°W-30°E], (b)
western (west of 0°E) and (c) eastern Sahel (east of 0°E). Vertical bars for the MMM indicate
multi-model STD, crosses indicate significant differences at 95% confidence level.



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

 2006-2100.



- 806 Figure 4: JAS horizontal wind (vectors) and moisture transport divergence (contour and
- shadings) at 925 hPa [kg/kg*s⁻¹]: (a) MMM of the CTL simulation and MMM differences
- between (b) 4K and (c) 4xCO2 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.



Figure 5: JAS global net radiative balance, defined at the top of the atmosphere as the
difference between the downward shortwave radiation and the sum of upward shortwave and
longwave radiation: (a) MMM of the CTL simulation and MMM differences between (b) 4K
and (c) 4xCO2 experiments and CTL simulation. In panel (a), black contours highlight the
separation between positive and negative values. Significant differences at 95% confidence
level are shaded. Note that the reference scale is not the same for the CTL simulation and the
sensitivity experiment responses.



- Figure 6: JAS (left) MSE and (right) DSE integrated in the 925-700 hPa layer: (top) MMM of
- the CTL simulation and MMM differences between (middle) 4K and (bottom) 4xCO2
- 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.





Figure 6: Continued.



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

simulations. Significant differences at 95% confidence level are displayed. Note that the

reference vectors are not the same for the CTL simulation and the sensitivity experiment

responses.



Figure 8: JAS zonal circulation averaged in [5°N, 20°N], displayed through zonal wind [m/s]

and vertical velocity [Pa/s] vectors in latitude-pressure plots. (a) MMM of the CTL

simulation and MMM differences between (b) 4K and (c) 4xCO2 experiments and CTL

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

same for the CTL simulation and the sensitivity experiment responses.



Figure 9: JAS velocity potential $[m^2/s]$ (contours and shadings) and divergent wind (vectors) at 200 hPa: (a) MMM of the CTL simulation and MMM differences between (b) 4K and (c) 4xCO2 experiments and CTL simulation. Divergent wind is computed as the gradient of the velocity potential. Vectors and shadings indicate differences significant at 95% confidence level. Note that reference scales and vectors are not the same for the CTL simulation and the sensitivity experiment responses.



Figure 10: JAS 925-700 hPa thickness [m], defined as the difference in geopotential height

between 700 and 925 hPa: (a) MMM of the CTL simulation and MMM differences between

(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

simulation and the sensitivity experiment responses.



- Figure 11: JAS meridional profile of (top) MSE and (bottom) DSE, integrated in the 925-700
- 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
- displayed. Note the differences in the y-axis values.
- 863



Figure 12: JAS zonal wind at 600 hPa [m]: (a) MMM of the CTL simulation and MMM
differences between (b) 4K and (c) 4xCO2 experiments and CTL simulation. Significant
differences at 95% confidence level are shaded. Note that the reference scale is not the same
for the CTL simulation and the sensitivity experiment responses.



Figure 13: Correlation between the JAS MMM of SRI and zonal wind at 600 hPa for the

period 1979-2008: a) CTL simulation, b) 4K and c) 4xCO2 experiments. Correlations

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



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