On the projected increase of Sahel rainfall during the late rainy season

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ABSTRACT: Thirteen CMIP5 models are used to analyse changes in climate over the West African monsoon region between the near future (2031–2070 under the RCP4.5 emission scenario), and a control period (1960–1999 under the historical emission scenario), with a focus on the late rainy season. The monsoon circulation is projected to strengthen and to shift northward leading to more rainfall during the Sahelian season. The results show an increase of the Rainfall amounts in September–October and a delay in the monsoon withdrawal. The increased moisture that fuels the rainfall anomalies is associated with an increase in moisture flux convergence and with local moisture recycling. The moisture transport dominates the water budget change in September while the local recycling is prominent in October. The delay in monsoon withdrawal, although expected from the increase in rainfall in September–October, is not strongly correlated with the size of the monthly anomalies.

KEY WORDS climate change; Sahel; CMIP5; delay

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1. Introduction

Monsoon circulations are linked to the seasonal cycle of solar heating and to differences in thermal inertia between land and ocean that establish a land-sea temperature and moist static energy gradient (Webster *et al.*, 1998; Nie *et al.*, 2010). They are associated with a global-scale persistent atmospheric overturning circulation (Trenberth *et al.*, 2000). Following the migration of the solar heating maximum, tropical rainy seasons occur in May–September in the Northern Hemisphere and in November–March in the Southern Hemisphere. During these periods, moist convection increases, becomes more organized and brings abundant precipitation, sustaining widespread rainfed agriculture.

Observations indicate that the onset of the West African monsoon (WAM) occurs at the end of June (Sultan and Janicot, 2000, 2003; Fontaine and Louvet, 2006) as an abrupt shift of the rainfall maximum from the Guinean coast to the Sahel, the ‘monsoon jump’ (Sultan and Janicot, 2003; Hagos and Cook, 2007). The dynamic of the onset has been linked to the heat low (Sultan and Janicot, 2003; Lavaysse *et al.*, 2009), the cold-tongue set up (Okumura and Xie, 2004; Caniaux *et al.*, 2011) and to complex interactions among convective processes in the InterTropical Convergence Zone (ITCZ), circulation associated with local topography and African easterly jet (AEJ) dynamics (Sultan and Janicot, 2003). The demise occurs between

September and November in West Africa (Liebmann *et al.*, 2012): first in the Sahel and then in Guinea. An early (late) demise has been associated with an anomalously strong (weak) North Atlantic subtropical high (NASH), which favours the increase (decrease) of the southerly moisture flux leading to an increase (decrease) of the moisture flux convergence over the Sahel (Zhang and Cook, 2014).

A positive rainfall trend from the nineties to present has been described as a ‘recovery’ of rainfall over the Sahel (Nicholson, 2005; Lebel and Ali, 2009) following the drought of the seventies and eighties. Recently Sanogo *et al.* (2015) have shown that the increase of rainfall over the Sahel during the period 1980–2010 is more prominent during the heart and the end of the season in August–September–October. In the context of climate change, a consensus is emerging to indicate similar changes. The most recent studies, based on multi-model analysis, show that the summer precipitation is projected to increase over the central Sahel and to decrease over the western Sahel (Monerie *et al.*, 2012; Biasutti, 2013; James *et al.*, 2015). The monsoon strengthening has been associated with an increase of the land-sea temperature gradient (as described by Haarsma *et al.*, 2005), of the cross-equatorial SST gradient in the tropical Atlantic (Hoerling *et al.*, 2006) and of the gradient between the extratropical Atlantic and the Tropics (Park *et al.*, 2015). This gradient would also be modified by anomalies in the eastern tropical Atlantic that reflect a change in the Atlantic circulation (Kröger *et al.*, 2005; Grist *et al.*, 2010; Servain *et al.*, 2014).

The more limited rainfall decrease over the western Sahel has been associated with an anomaly of subsidence

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Table 1. Names, references and grid resolutions of the 13 CMIP5 models used in this study.

Models	References	Institution	Resolution
canesm2	Chylek <i>et al.</i> (2011)	Canadian Centre for Climate Modelling and Analysis (Canada)	128 × 64 × 22
cnrm_cm5	Voldoire <i>et al.</i> (2013)	Centre National de Recherches Météorologiques (France)	256 × 128 × 17
csiro_mk3_6_0	Rotstayn <i>et al.</i> (2010)	CSIRO Atmospheric Research (Australia)	192 × 93 × 18
giss_e2_r	Shindell <i>et al.</i> (2013)	Goddard Institute for Space Studies (USA)	144 × 89 × 17
hadgem2_es	Collins <i>et al.</i> (2011)	Met Office (UK)	192 × 144 × 17
inmcm4	Volodin <i>et al.</i> (2010)	Institute for Numerical Mathematics (Russia)	180 × 120 × 17
ipsl_cm5a_lr	Dufresne <i>et al.</i> (2013)	Institut Pierre-Simon Laplace (France)	96 × 96 × 17
mri_cgcm3	Yukimoto <i>et al.</i> (2012)	Meteorological Research Institute (Japan)	320 × 160 × 23
bcc_csm1-1	Xiao-Ge <i>et al.</i> (2012)	Beijing Climate Center (China)	128 × 64 × 17
bnu_esm	More informations on the websites: http://esg.bnu.edu.cn/BNU_ESM_webs/htmls/index.html and http://esg.bnu.edu.cn/BNU_ESM_webs/htmls/index.html	Beijing Normal University (China)	128 × 64 × 17
gfdl_esm2m	Dunne <i>et al.</i> (2012)	Geophysical Fluid Dynamics Laboratory (USA)	144 × 91 × 17
miroc_esm	Watanabe <i>et al.</i> (2011)	Center for Climate System Research Institute (Japan)	128 × 64 × 35
noresm1_m	Bentsen <i>et al.</i> (2012)	The Norwegian Climate Centre (Norway)	144 × 96 × 17

and a strengthening of the AEJ (Monerie *et al.*, 2012; James *et al.*, 2015).

Climate change has also been shown to impact the WAM during the demise months, in SO (September–October): the entire Sahel is projected to become wetter in response to increase greenhouse gases (Biasutti and Sobel, 2009; Wang and Alo, 2012; Biasutti, 2013; Seth *et al.*, 2013). Using general circulation model (GCM) and regional climate models (RCMs), Vizy and Cook (2012) and Vizy *et al.* (2013) have found that the increase in seasonal total Sahel rainfall follows from more intense rainfall, and no appreciable change in the length of the rainy season.

The aim of this study is to address two main questions: (1) Is a consensus on SO Sahel rainfall emerging around the middle of this century in a reduced-emission scenario? (2) What are the mechanisms responsible for the changes?

This article is organized as follows. Data and methods are presented in Section 2. This study focuses on rainfall anomalies in Section 3 and on the associated moisture flux changes in Section 4. Section 5 allows an analysis of the changes occurring specifically in September and October. We address the question of the demise date in Section 6. Section 7 concludes.

2. Data and method

2.1. Climate change analysis

We select 13 CMIP5 model outputs downloaded from the Program For Climate Model Diagnosis and Intercomparison (PCMDI) server (<http://pcmdi9.llnl.gov/esgf-web-fe/>) and used in the fifth phase of the Coupled Model Inter-comparison Project CMIP5 (Taylor *et al.*, 2012): names, references and resolutions are given in Table 1.

The present climate is defined as mean of the historical simulation over the period 1960–1999 (and is noted ‘CTRL’ for the control simulation), the future climate is defined as the Representative Concentration Pathway (RCP) 4.5 simulation from 2031 to 2070 (and is noted ‘FTR’ for the future simulation). RCP4.5 is a medium-low emission scenario reaching about 4.5 W m^{-2} ($\sim 540 \text{ ppm CO}_2$) in year 2100 (Meinshausen *et al.*, 2011; Thomson *et al.*, 2011). According to Rogelj *et al.* (2012), the RCP4.5 is comparable to the Special Report on Emissions Scenarios (SRES) B1 with similar CO_2 concentrations; the ensemble median temperature rises quickly until mid-century, more slowly afterwards and exceeds $+2^\circ\text{C}$ warming after 2100. Climate change effects are computed as the differences between the mean FTR and the mean CTRL climates. As the time evolution of monsoonal rainfall is linked to Sea Surface Temperature (SST) variability, including the Atlantic multidecadal oscillation, the Pacific decadal oscillation, the interdecadal Pacific oscillation and SSTs in the Gulf of Guinea, the results may be influenced by the SST decadal variability in a warming world. To avoid this bias, this study uses 40-years periods and a multimodel ensemble. Decadal and multidecadal variability are thereby in partly smoothed. The multidecadal periodicity of the SSTs (such as the interdecadal Pacific oscillation) can impact the Sahel rainfall low-frequency variability (Villamayor and Mohino, 2015), but the multimodel average ensures that this variability is also smoothed because occurring on different timings for each model.

The multimodel analysis minimizes biases (Lambert and Boer, 2001; Phillips and Gleckler, 2006; Randall *et al.*, 2007; Gleckler *et al.*, 2008; Pincus *et al.*, 2008; Reichler and Kim, 2008; Pierce *et al.*, 2009), but several errors common to models can remain, such as a poor cold tongue simulation (Richter and Xie, 2008) in the Gulf of Guinea. However, this method appears to be better than the choice

of a ‘best model’ (Knutti *et al.*, 2010), because model selection is never evident and depends on the scope of each study (Santer *et al.*, 2009).

The multimodel approach assumes that the models used are independent, avoiding any over-representation of one type of model configuration. Thus, we chose only one model per climate centre (13 models) and one simulation per model to give the same weight to each climate centre and to maximize model independency (as suggested by Knutti *et al.*, 2010; Masson and Knutti, 2011; Pennell and Reichler, 2011). All fields are interpolated on the same $2.5 \times 2.5^\circ$ resolution. We complement the ‘FTR’ minus ‘CTRL’ difference maps with maps of occurrences based on the ‘one model, one vote’ concept proposed by Santer *et al.* (2009) and used by Fontaine *et al.* (2011) and Monerie *et al.* (2012, 2013). This metric indicates whether the signal is due to several outlier models or to the majority of models.

2.2. Moisture flux decomposition

In this article, the surface moisture fluxes are broken into its mean circulation dynamics (MCD) and the thermodynamic (TH) components:

$$\delta_{TH} = \overline{\vec{u}_{CTRL} \delta q}$$

$$\delta_{MCD} = \overline{\delta \vec{u} q_{CTRL}}$$

where,

$$\delta(\cdot) = (\cdot)_{FTR} - (\cdot)_{CTRL}$$

is the RCP-historical change, bars are the monthly means and subscripts FTR and $CTRL$ indicate the FTR and CTRL period values of the arbitrary quantity in parentheses. The decomposition is performed for the zonal (u) and meridional (v) components of the wind ($\vec{u}$).

This decomposition is inspired by Seager *et al.* (2010), but it is limited to low-level fields, where moisture flux is greatest.

2.3. Demise date

The withdrawal of the monsoon is more abrupt than the set-up which is marked by active and break phases (Louvet *et al.*, 2003). Various methods are employed to detect onset/demise dates. Here we follow the methodology developed by Liebmann *et al.* (2012) based on the difference between the daily precipitation and the long term annual mean daily average.

For each year and grid point, the precipitation annual mean is computed and subtracted from the raw daily value of precipitation to give an accumulation anomaly (A) as a function of calendar day

$$A(\text{days}) = \sum_{n=1}^{\text{days}} \left(R(n) - (\bar{R}) \right)$$

where $R(n)$ is the daily precipitation and $\bar{R}$ is the annual average. The day at which the accumulation anomaly

Figure 1. Rainfall amounts evolutions (mm day^{-1}) from January to December averaged over (a) the Sahel ($10^\circ\text{W}–10^\circ\text{E}$; $10^\circ–20^\circ\text{N}$) (the two domains are represented within Figure 2(a)). The mean CTRL climate is represented by a continuous black line (right abscissa) and the FTR minus CTRL differences by the bars for precipitation (grey bars) and evaporation (white bars) (left abscissa). The FTR–CTRL precipitation minus evaporation ($P-E$) result is represented with discontinuous lines and stars (left abscissa).

reaches its maximum is the day of demise of the rainy season. This method is more appropriate than the use of a simple threshold because there are large variations in the rainfall values from one model to another.

3. Rainfall amount and atmospheric circulation changes in September–October

Figure 1 displays the monthly mean rainfall climatology simulated for the Sahel [$10^\circ\text{W}–10^\circ\text{E}$; $10^\circ–20^\circ\text{N}$] (see the box in Figure 2(a)) in the CTRL simulation (continuous black line), along with the precipitation, the evaporation, and precipitation minus evaporation ($P-E$) differences between the FTR and CTRL periods (grey bars, white bars and stars). In the FTR period, the Sahel is projected to become wetter than CTRL from July to October: during the core of the rainy season and during the demise months. This is associated with an increase in moisture flux convergence (positive values of the $P-E$ change). In October, the large evaporation anomalies suggest that local water recycling could play a key role in producing a wetter late season. More precipitation during the last months of the monsoon is consistent with the literature as reported by Kitoh *et al.* (2013) and Lee and Wang (2014) for the global monsoon and by Biasutti and Sobel (2009), Biasutti (2013) and Seth *et al.* (2013) for the WAM. The increase of rainfall at the end of the northern tropical rainy season is the focus of this work and may denote a delay in the monsoon withdrawal.

The increase of precipitation in September–October (SO) is significant over the entire Sahel (Figure 2(a)) and robust (it is reproduced by a large majority of the models, up to 10 models out of 13; Figure 2(b)). A maximum in precipitation anomaly is located around the Greenwich meridian (box in Figure 2(a); $10^\circ\text{W}–10^\circ\text{E}/10^\circ–20^\circ\text{N}$). This rainfall increase is over and northward of the mean

Figure 2. (a) Mean SO rainfall amounts (mm day^{-1}) over the CTRL period (grey lines) and for the FTR-CTRL differences (shadings). The values judged significant with a Student's t -test at $p=0.05$ ($p=0.10$) are marked by a plus (cross). (b) Occurrence map: a value of +11 or +12 (–11 or –12) means that 11 or 12 models are in accordance with a positive (negative) change in multimodel differences. Sixteen CMIP5 models have been used.

rain band in the CTRL simulation (continuous grey lines) and represents a strengthening and a northward shift of the monsoon system.

This is consistent with the anomalies in low-level (925–850 hPa) winds, which are located between 10°N and 20°N , north of the climatological southwesterlies converging into the Sahel (Figure 3(a)). It is well known that a northward position of the AEJ and a strengthening of the Tropical Easterly Jet (TEJ) are related to wet years (Grist and Nicholson, 2001). At mid-level, there is no significant change of the AEJ speed and position (located between 700 and 600 hPa and from 5° to 15°N in the mean CTRL period). The TEJ seems to move southward (see the negative anomalies between 5°S and 5°N at 250 hPa) but it is not well defined in the CTRL period. A significant upper-level easterly anomaly at 25°N could be linked to the northward shift of the jet shown by Yin (2005).

The main changes of the zonal winds are thus the strengthening of the low-level south-westerlies and a weakening of the subtropical westerly jet; both are reproduced by at least ten models (Figure 3(c)).

Along with these winds, the monsoon cell moves northward and West Africa experiences more subsidence from 5° to 15°N and more ascent north of 15°N and south of 25°N (Figure 3(b)). The negative values of the omega anomalies are significant (and denote by convention more ascending motions) and robust (Figure 3(d)). Deep ascending anomalies extending from 925 to 200 hPa are co-located with the increase in rainfall up to 20°N . At

low-level, the omega anomalies indicate a northward shift of the monsoon flow pushing the confluence boundary and the shallow dry ascent to 23°N .

The increase in rainfall amounts in SO is associated with the strengthening and northward shift of the monsoon circulation. What is the source of moisture for the additional rainfall? This question is explored in the next section.

4. Changes in moisture supply

Evaporation anomalies are positive year round (Figure 1; white bars) but are largest during the late months of the rainy season because evaporation anomalies are in large part a consequence of the accumulated rainfall anomalies. Evaporation is moisture-rather than energy-limited in the semi-arid Sahel (Cook *et al.*, 2014). The $P - E$ change (Figure 1; stars) indicates an increase of moisture convergence over the Sahel during the rainy season, with largest values in August and September.

Kitoh *et al.* (2013) showed that projections of rainfall surplus in the global monsoon area are mainly due to increases in specific humidity, which produces an increase in moisture flux convergence, even in the face of weakening circulations. This explanation does not seem to hold for the WAM, as the simulations project large changes in the regional circulation and indeed a strengthening of the monsoon flow and of deep ascent.

Figure 4 shows the maps of wind and sea level pressure (Figure 4(a)), moisture and moisture flux at 825 hPa (Figure 4(b)) for the mean CTRL SO period. In SO, the main source of moisture for the monsoon rain is the Mediterranean Sea and the North Atlantic Ocean with northerlies (Figure 4(a)) converging moisture over West Africa (Figure 4(b)).

Figure 4(c) and (d) show SO FTR-CTRL differences of sea-level pressure and winds (Figure 4(c)), and specific humidity and moisture flux (Figure 4(d)) at 825 hPa. The Sea Level Pressure (SLP) anomalies show anticyclonic anomalies over mid-latitude continental Europe and the western Mediterranean region, corresponding to a northward shift of the NASH (Figure 4(c) and (e)). Over Africa, cyclonic anomalies indicate a deeper heat low (Figure 4(c)). The anomalous pressure gradients favour significant increases in the northerlies and southerlies converging in the northern Sahel (Figure 4(c)). As expected from the global temperature increase, the specific humidity is higher over the entire domain (Figure 4(d)). The anomalous moisture flux converges moisture over the Sahel mainly from the eastern Mediterranean Basin. These changes are significant and reproduced by a large majority of models (Figure 4(f) and (g)).

Examining the components of the moisture fluxes shows how a strengthening of the winds and increases in specific humidity affect the $P - E$ changes. The surface moisture flux anomalies are broken down into a component due to changes in monthly means wind (the MCD) and a component due to the monthly mean changes in surface specific

Figure 3. Latitude-height cross section (average from 10°W–10°E) of (a) zonal winds (m s^{-1}) and (b) omega (hPa). The mean CTRL is displayed with grey contours and the FTR–CTRL change with shadings. The values judged significant with a Student's t -test at $p = 0.05$ are marked by a plus. Maps of occurrences (number of models) of the FTR minus CTRL differences of (c) zonal wind and (d) omega.

humidity (TH) at 825 hPa. The results are displayed in Figure 5.

The dynamic and TH components are of the same magnitude (Figure 5(a) and (b)). The moisture flux change associated with the TH is more important in northern Sahel (at 20°N) and MCD anomalies are associated with more inflow from the tropical Atlantic south of the Sahel (12°N) (Figure 5(a) and (b)). The northward shift of the monsoon is therefore dominated by the change in dynamic but a large part of the moisture increase is due to the strengthening of the northerlies. The strengthening of the moisture flux associated with the MCD and TH is robust over the Sahel (Figure 5(c) and (d)).

We can thus draw the conclusion that the late monsoon season is projected to be wetter (Figure 2(a)) due to increased moisture fluxes (Figure 4(d)) associated with the global temperature rising and to a northward shift of the monsoon system and of the North Atlantic high (Figure 4(b)). This is a classic view of the climate change impact on the WAM that can also be drawn for the JAS

time-period (Maynard *et al.*, 2002; Monerie *et al.*, 2012, 2013; James *et al.*, 2015, among others, and suggested in Figure 1). The aim of this study is, however, to analyse the changes specific to the late monsoon season. Figure 1 shows stronger increases of moisture flux convergence ($P - E$) in September than in October suggesting larger role for dynamics in September. Such differences are analysed in the next section.

5. September and October changes

We compute the S–JJASO and O–JJASO anomalies to stress the particularity of each month with respect to the extended rainy season. We define

$$M - \text{JJASO} = \text{FTR}_{(M-\text{JJASO})} - \text{CTRL}_{(M-\text{JJASO})} \quad (1)$$

where M is the month we are looking for (i.e. September or October), $\text{FTR}_{(M-\text{JJASO})}$ and $\text{CTRL}_{(M-\text{JJASO})}$ are the seasonal means, respectively, for the FTR and CTRL periods.

Figure 4. SO mean of (a) sea level pressure (Pa) and wind (m s⁻¹), (b) specific humidity (g kg⁻¹) and moisture flux (g kg⁻¹ × m s⁻¹) at 825 hPa, SO FTR-CTRL differences of (c) sea level pressure and winds, (d) specific humidity and moisture flux at 825 hPa. SLP values are shown with shadings and superimposed grey plus when the differences are significant at $p = 0.05$. For more clarity, only the significant moisture flux differences are represented (vectors). Maps of occurrences (number of models) of the FTR minus CTRL differences of (e) SLP, (f) the zonal component of the moisture flux and (g) the meridional component of the moisture flux.

Figure 5. (a) Mean circulation dynamic (MCD) and (b) thermodynamic (TH) in SO (g kg^{-1}) and at 825 hPa. Maps of occurrences (number of models) for the meridional component of the (c) MCD and (d) TH.

Please note that these anomalies are not directly comparable to the results displayed in Figures 2–5.

Figure 6 displays the S–JJASO (first column) and the O–JJASO (second column) anomalies of SLP, 825 hPa moisture fluxes and rainfall. In September, the anomaly of SLP indicates a deeper heat low centred at 10°E and North of the Sahel (Figure 6(a)), although the anomaly is not significant across the ensemble. Rainfall increases in association with the anomalous northerly moisture flux (Figure 6(a) and (e)). The O–JJASO anomaly exhibits an increase in SLP over Europe and the Mediterranean and extending over the Sahara. In association with it, there is an increase in moisture flux from the eastern Mediterranean towards the Sahel (Figure 6(b) and (f)). Nonetheless, the strongest moisture flux anomaly is south of 20°N indicating that the moisture flux change is mainly due to local changes of moisture over the Sahel. This is consistent with Figure 1, which clearly indicates that local evaporation is the dominant contribution to the water budget change in October. Thus, rainfall increases just south of the northerly moisture flux increase (Figure 6(b))

6. The withdrawal date of the WAM

The increase in rainfall during September and October may denote a change in the demise date of the rainy season. As described in Section 2.3, we define the end of the monsoon as the date when daily precipitation systematically drops below its local annual mean, as described by Liebmann

et al. (2012). A demise date is computed for each grid point and each model. The mean demise date of the Sahel is then obtained as the average of each demise date for each grid point covering the central Sahel (defined as in Figure 2(a): 10°W – 10°E ; 10° – 20°N). The sensitivity to such choice is checked by using two other boxes, namely the western Sahel (15°W – 0°E ; 10° – 20°N) and the eastern Sahel (0° – 25°E ; 10° – 20°N). For the CTRL period, the mean demise dates over the western, central and eastern Sahel are 4 October, 30 September and 22 September. These demise dates are consistent with the study of Liebmann *et al.* (2012) who found that the demise occurs between September and November in West Africa, and late over western than central Sahel.

The FTR minus CTRL differences are shown in Figure 7(a). Over the entire Sahel, the demise date are projected to occur several days later, denoting a delay of the rainy season withdrawal. In Figure 7(b), we plot the inter-model spread in the anomalies in demise date as a function of the SO mean change for the western, the central and the eastern Sahel. We find a weak dependence between the two quantities and a linear relationship is statistically significant only for the western Sahel. The weak relationship is especially due to anomalies in the *miroc_esm*, which is an outlier model for the region (Figure 7(b)). This model does simulate an increase in precipitation in SO but the increase during JAS is a lot stronger (the increase reaches 3 mm day^{-1} , i.e. ten times the multimodel change) and the shape of the accumulation

Figure 6. Moisture flux ($\text{g kg}^{-1} \times \text{m s}^{-1}$) at 825 hPa, sea level pressure (Pa) and rainfall (mm day^{-1}) for the FTR–CTRL (a) S-JJASO differences and (b) O-JJASO differences. SLP values are shown with shadings and superimposed grey plus when the differences are significant at $p=0.05$. For more clarity, only the significant moisture flux differences are represented (vectors). The rainfall differences are displayed with continuous lines for positive values and discontinuous lines for negative values. Maps of occurrences (number of models) of the FTR minus CTRL differences of (c) S-JJASO SLP, (d) O-JJASO SLP, (e) S-JJASO meridian component of the moisture flux and (f) O-JJASO meridian component of the moisture flux.

anomaly is not changed (not shown). A large majority of the models simulate a delayed demise date along with the increase in SO rainfall.

For Zhang and Cook (2014), a late demise date is associated with a weak NASH and increased southerlies while we found more rainfall during the demise months because of a local increase of moisture convergence and northerlies. These two results are not excluding each other because these mechanisms do not occur on the same time-scale (daily in the study of Zhang and Cook, 2014 and monthly here). Moreover, our definition of the demise date differs: we chose a measure that takes into account the entire rainfall season, while they use a fixed threshold of 2 mm day^{-1} .

Our choice of how to define the demise of the rainy season has the advantage of being compatible with the coarse-resolution of our rainfall data. Yet, it is not the definition that is used on the ground: because the timing of the rainy season is tracked for agricultural purposes, the demise of the rainy season is typically defined as the cessation of the growing season (when soil moisture is so depleted that it cannot support crop growth). Following Lodoun *et al.* (2013), we compute the cessation date of the growing season for the CTRL and FTR periods. In the latter case, we scale the daily potential evapotranspiration to take into account the effect of higher temperature (in accordance with Scheff and Frierson, 2014, we imposed

Figure 7. (a) Mean FTR–CTRL demise date change (days) for the western (15°W–0°E; 10°–20°N), central (10°W–10°E; 10°–20°N) and eastern (0°–25°E; 10°–20°N) Sahel. The box plot indicates the demise date change median (line), 25th and 75th percentile. (b) Demise date change (days) in function of the mean change in precipitation in SO (mm day⁻¹) for the western (pluses), central (stars) and eastern (diamonds) Sahel. Each symbol represents one of the 13 CMIP5 models used.

an increase of 0.3 mm day⁻¹). By this definition, the cessation date of the growing season is found to occur earlier (but with large scatter across models). This is due to a FTR–CTRL potential evapotranspiration increase that is greater or equal to the increase in precipitation.

By this calculation, the delay of the climatological demise date has thus no impact on the cessation date of the growing season and thus on the agricultural activity. Yet, this result is in contradiction with the work of Cook and Vizy (2012), who found a week-long delay in the cessation of the growing season. The discrepancy might result from the different model sensitivities (Cook and Vizy (2012) used one RCM forced with CMIP3 climate change), or from our rather crude representation of potential evapotranspiration changes. Moreover, the start and end of the growing season depend on the specific sensitivity of each crop to temperature, moisture and length of day (for example see Guan *et al.*, 2015). Thus a lot more work is needed to untangle the relationship between seasonal rainfall anomalies, the demise of the monsoon and the cessation of the agricultural rainy season.

7. Summary and discussion

We study the projected changes in rainfall and atmospheric circulation in West Africa during the demise months (SO). We focus on the multimodel mean but we choose only one version of each available CMIP5 model to maximize the data independency (as suggested by Masson and Knutti, 2011; Pennell and Reichler, 2011; Knutti *et al.*, 2013). Robustness is measured by the number of models that agree with the sign of the multimodel response. Even for a weaker forcing (a more benign scenario than chosen in most previous CMIP5 studies and looking at the near future), we confirm the future strengthening of the WAM in SO (as described by Biasutti and Sobel, 2009; Wang and Alo, 2012; Biasutti, 2013; Seth *et al.*, 2013).

The mechanisms by which the late-season rainfall increases under global warming differ between the northern and southern Sahel and between September and October. Changes in the monthly mean circulation are especially important for the northern Sahel and during September. Changes in the overall moisture content of the lower troposphere and in local precipitation recycling and evaporation are more important for the southern Sahel and during October. Mid-latitude changes and the increase of the northerlies from the Mediterranean to the Sahel can explain a small part of the rainfall change in SO.

The late season in SO is characterized by a delay of the demise date in the future. It occurs between the middle of September to the beginning of October. The demise starts first in eastern Sahel, then central and western Sahel. There is a weak linear relationship between the delay of the demise date and the precipitation anomalies change in SO. However, for most of the models, a late demise date is associated with higher SO rainfall amounts.

The simulations of the CMIP5 atmosphere–ocean–general circulation model (AOGCM) are marred by several biases due to the models' coarse resolution and a drastic underestimation of regional orography. The present results should be confirmed with RCMs and global GCM simulations using finer grids. This will allow better simulations, for example, of the Mediterranean moisture source but also of wind changes due to small scale processes and better representation of the strong gradients in atmosphere–land fluxes and moisture recycling. It is however crucial to use consistent physical packages between the RCM and the forcing GCM (Saini *et al.*, 2015). Otherwise the downscaled field will not be the result of the forcing field with a higher resolution but mainly to a change of the parameterization.

Furthermore, the CMIP5 AOGCM simulations used in this study do not take into account the vegetation and land use evolutions. The precipitation–vegetation feedback could have a considerable impact on the sub-Saharan

region (Douville *et al.*, 2001; Paeth and Thamm, 2007; Paeth *et al.*, 2009; Wang and Alo, 2012). Adding land use and vegetation evolutions could allow more accurate future projections and may lead to a stronger increase of rainfall amounts during the last months of the monsoon period through positive feedbacks: more rainfall amounts are associated with more vegetation cover and thus more water recycling.

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