Recent trends in the Central and Western Sahel rainfall regime (1990–2007)

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S U M M A R Y

One motivation for setting up the CATCH (Couplage de l’Atmosphère Tropicale et du Cycle Hydrologique) project at the end of the 1990s, was to contribute to documenting the Sahelian rainfall variability at the interannual scale and to provide a fine monitoring of possible long-term trends of the rainfall regime. This paper is a first attempt at characterising the Sahelian rainfall regime of the two last decades (1990–2007) by comparison to the rainfall regime of the previous decades, namely the 20-year wet period 1950–1969 and the 20-year dry period 1970–1989. While the rainfall deficit remained unabated in the Western Sahel (1990–2007 mean equal to the 1970–1989 mean, both being lower than the 1950–1969 mean), the Central Sahel progressively recorded wetter years from the end of the 1990s, but this recovery is limited (1990–2007 average larger by 10% than the 1970–1989 average, but still lower than the 1950–1989 average). There are also significant differences between the Western Sahel and the Central Sahel when looking at the interannual variability pattern and at the seasonal cycle. The low-frequency rainfall patterns are similar between the Western Sahel and the Central Sahel, but the interannual year-to-year variability is weakly related to each other. In the Central Sahel, the major modification of the seasonal cycle in the most recent decades was the disappearance of the well marked August peak observed during the wet period. In the Western Sahel the rainfall deficit is more or less evenly distributed all along the rainy season. The second part of the paper makes use of the CATCH-Niger recording rain gauge network in order to compare several ways of defining rainy events. The statistical properties of these various populations of rainy events are compared. It is shown that a simple CPP model allows for retrieving the statistical characteristics of point rainy events from daily rainfall series. It is also confirmed that in this area, the interannual rainfall variability is primarily linked to the year-to-year fluctuation of the number of large mesoscale rainfall events.

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Introduction

The generalised drought that struck West Africa during the 1970s and 1980s was the most significant climatic event at the regional scale during the 20th century (AMMA ISSC, 2005), prompting the international scientific community to launch the AMMA (African monsoon multidisciplinary analyses) program, a long-term integrated research project (Redelsperger et al., 2006) on the causes and impacts of this “anomaly”. Le Barbé et al. (2002) showed that the annual rainfall deficit of the 1970–1989 period, as compared to the rainfall of the previous 1950–1969 period, covered the whole region (6,000,000 km\textsuperscript{2}), with an average of about 190 mm (see Fig. 1 of Lebel et al., this issue). This means that in the northern part of the domain where the 1950–1969 annual rainfall was in the order of 300–400 mm, the relative deficit of the dry 1970–1989 period amounted in average to 50–75% during 20 consecutive years. The consequences of this drought were devastating for the region, causing famines and migrations, and playing a significant role in the low economic growth of the region (see e.g. Davidson et al., 2003; ECOWAS-SWAC, 2006; Stige et al., 2006).

Superimposed on this decadal signal, marked interannual variations are another key characteristic of the rainfall regime of West Africa. For the Gulf of Guinea coastal region, the interannual variability dominates the rainfall time series for most of the last century (e.g., Rowell et al., 1995; Ward, 1998); the further north across the Soudan and Sahel zones, the larger is the predominance of the decadal signal for the second half of the 20th century.

These variations are determined by those of the West African monsoon (WAM), which is characterised by several key approximately zonal flows that are established in association with the meridional heating contrasts and associated direct circulations (Fig. 1). The African easterly jet (AEJ) is located in the region of strong low-level potential temperature gradients between the Sahara and the Guinea Coast (see $\theta_e$ profiles in Fig. 1). At low-levels south-westerlies from the Atlantic provide most of the moisture for the WAM while polewars of this north-easterlies advec relatively drier Saharan air into the rainy region.

The ITCZ (inter-tropical convergence zone) is the primary factor controlling directly the rainfall over West Africa, due to the associated large-scale convergence. In boreal winter the ITCZ is located around 5°S on the Tropical Atlantic and the continent is dry. The ITCZ then moves to the north, following the northward migration of the maximum of received solar radiation energy. It reaches its most northern position in August between 10°N and 12°N before retreating to the south. The ensuing typical seasonal rainfall cycle over the Sahel, north to 11°N, is a 3–6 month rainy season with a peak of intensity in August.

While, from a large-scale perspective the WAM can be described in terms of the annual march of the ITCZ and its associated regional circulations, smaller-scale structures are pivotal for rainfall production. At the regional scale, Grist and Nicholson (2001) have shown that a farther north than average location of the AEJ is associated with wet years. Since the ITCZ keeps a relatively stable position at the surface from year to year they infer that the interannual rainfall variability in the Sahel is critically controlled by the position of the AEJ, rather than its intensity. The African easterly waves (AEWs) developing in the AEJ (e.g. Hall et al., 2006; Kiladis et al., 2006), play an important role in the development of large organised mesoscale convective systems (MCs), responsible for most of rainfall over the region. The way in which the MCS interact with the surface and the large-scale environment including the jets and monsoon layer winds (e.g. Redelsperger et al., 2002) is determinant for the effective production of rainfall.

It stems from this brief description that the WAM dynamics is basically a response to the contrast in temperature and humidity between the continent and the Tropical Atlantic, the variations of the surface conditions of these two areas are thus both susceptible to largely contribute to the rainfall variability at various temporal scales.

It is generally accepted that sea surface temperatures have a major role in determining interannual-to-decadal variability of the West African monsoon; Hastenrath (1990), Ward (1998), Gianinni et al. (2003) for instance, point to the multidecadal decreasing trend of WAM rainfall being closely linked to the change of the north–south interhemispheric gradient of SST and related changes in the atmospheric circulation. The role of the land change is more difficult to quantify, even though Charney (1975) pointed out very early that vegetation degradation over the Sahel might induce a long-term inhibiting feedback on rainfall. Since both the long-term warming in the Indian Ocean – which is the main component having affected the north–south interhemispheric gradient of SST – and the vegetation clearing are still effective, an expected consequence would be for a continuation of the rainfall deficit at the decadal scale. Recent works on the subject led to divergent conclusions: while Ozer et al. (2003) diagnosed that the Sahelian drought ended during the 1990s, Nicholson (2005) conclude that a rainfall recovery did occur in the northern Sahel (north to 16°N) over 1998–2003 but that this recovery was limited further south.

In this context where both the recent rainfall trends and the possible causes of the decadal variability are subject to discussion, the aim of this paper is to provide a factual update on the recent rainfall observations available to the scientific community over the Sahel and to analyse what they teach us about recent regional rainfall trends. Complementing the earlier works cited above, we are now able to look at a period of length similar to the reference wet (1950–1969) and dry (1970–1989) periods used in previous studies over the region (i.e., Le Barbé et al., 2002) and thus to compare the present rainfall regime to those of these periods. One limitation for this type of study is the relatively small number of long-term rainfall series readily available to the scientific community. The operational networks (Ali and Lebel, 2009) of the nine Sahelian countries total about 650 daily-reading rain gauges, but only 266 of these stations (R266 network) have less than 20% of missing data on the whole 1950–2006 period. A subset of 96 stations has operated continuously over the same period, forming the R96 network. These two networks are shown in Fig. 2, along with the various study areas used in the paper to characterise the interannual variability, the seasonal cycle and the daily rainfall distributions of the recent period (1990–2007) as compared to those of the 1950–1969 wet period, and of the 1970–1989 dry period. Three large 10° (in longitude) × 5° (in latitude) boxes will be used first to obtain a global view of the longitudinal gradients of the Sahelian rainfall at the interannual scale (the western box is in fact 8° wide in longitude, while e 5°) and moist static energy (\( \theta_e \)). North to 10°N \( \theta_e \) starts to decrease while \( \theta \) continues to increase, due to the drying of the air mass north to the core of the ITCZ (after the AMMA International Science Plan, 2005, adapted from Parker et al., 2005).

Fig. 1. Schematic of the atmospheric circulation in the West African monsoon system during the boreal summer. Closed solid lines represent the streamlines at the level of the African easterly jet (AEJ) around 600 hPa. Grey shading represents peak rainfall and yellow shading indicates the location of the Saharan Air Layer (SAL). Below are given the typical corresponding meridional variations in atmospheric boundary layer potential temperature (\( \theta \)) and moist static energy (\( \theta_e \)). North to 10°N \( \theta_e \) starts to decrease while \( \theta \) continues to increase, due to the drying of the air mass north to the core of the ITCZ (after the AMMA International Science Plan, 2005, adapted from Parker et al., 2005).

since the 20°–18°W area belongs to the ocean). The study will then focus on two 5° × 5° boxes assumed to be representative of the Western Sahel rainfall regime, on the one hand, and of the Central Sahel rainfall regime, on the other hand.

In addition we will use the high space-time resolution data set provided by the AMMA-CATCH-Niger (ACN) network over the shaded rectangle of Fig. 2 (see Cappelaere et al., this issue, regarding the instrumentation of the ACN area). This will allow looking at the rainfall regime at the rain event scale for the recent period 1950–2007.

A global view of the Sahelian rainfall over 1990–2007

There are relatively few studies that looked at the 1970–1990 drought in term of changes in the seasonal cycle and in the rain event distribution. Focusing on a 15° (10°W–5°E) × 11° (5°–16°N) window, Le Barbé et al., 2002 showed that, with respect to the 20-year wet period 1950–1969 (P1), the 20-year dry period 1970–1989 (P2) was characterised by a smaller number of rain events of roughly unchanged intensity and that this decrease was especially pronounced for the core of the rainy season.

By comparison to the previous periods, one major trend of the recent period P3 (1990–2007) is the growing contrast in the regional pattern of the Sahelian annual rain fields as illustrated by Fig. 3. In this figure a few isohyets were selected for purpose of comparison between the three periods: it clearly appears that while the Western Sahel is still suffering from a pronounced rain deficit, the eastern Sahel is recovering. In order to analyse this phenomenon in time, a rainfall anomaly index is computed with reference to the 1950–1989 period. Let \( P^{(k)} \) and \( s^{(k)} \) be respectively the average and the standard deviation of the annual rainfall over region \( k (k = 1, 3) \) and the period 1950–1989, then the rainfall anomaly of year \( j (j = 1950, 2007) \) is computed as:

\[
I_k(j) = \frac{P_k(j) - P^{(k)}}{s^{(k)}}(j)
\]  

(1)

This means that the rainfall anomaly series has a mean of 0 and a standard deviation of 1 over the reference period 1950–1989, and not over the whole period 1950–2007. This allows for a more meaningful comparison of the behaviour of each sub-region for the recent period \( P_3 \) (1990–2007), these sub-regions being defined as: \( k = 1: [11°–16°N; 20°–10°W]; k = 2: [11°–16°N; 10°W–0°]; k = 3: [11°–16°N; 0°–10°E]. \) Table 1 gives the main statistical parameters of each of these anomaly series for period \( P_3 \), while the rainfall anomaly time series are shown as insets in Fig. 3.

The difference of the mean of the anomalies between box 1 and box 3 is very significant, the average of box 3 being larger than the average +1 standard deviation of box 1. The averages of box 1 and box 2 are also significantly different (level 5% of a classical Student test). Another point to notice is that the western box time series displays smaller interannual variability over the recent years, as compared to the central and eastern boxes. If one retains the criterion set by Ali and Lebel, 2009 for a year to be deemed significantly dry when the anomaly index is below −0.5, then during the recent period \( P_3 \), the western box had 13 dry years out of a total of 18, while the central box had 10 and the eastern box only 7. Over the same period, the western box had not a single significantly wet year (anomaly larger than +0.5), while the central had 3 and the eastern box had 4 (1994, 1999, 2001, 2007).

### Western versus Central Sahel

While in some studies “Central Sahel” refers to a median position in latitude (typically 13°–15°N) this terminology refers here to a median position in longitude, between say 5°W and 10°E. The aim of this section is to analyse the seasonal cycle over two regions displaying a different behaviour over the recent years. This is done by selecting the two 5° × 5° boxes shown in Fig. 2. Reducing the length of the boxes in longitude from 10° to 5° aims at retaining more homogeneous sub-regions, given the west–east gradient observed above in term of rainfall recovery over the most recent period.

### Decadal trends and interannual variability

As shown in Fig. 4, the 15–10°W box is definitely dryer over period \( P_3 \) than the 0°–5°W box. This is in clear contrast with period
Determination coefficients between the three anomaly series for period $P_1$ (1990–2007). The mean ($m_{P_i}$) and standard deviation ($s_{P_i}$) of each series are also given in the first line of the table.

<table>
<thead>
<tr>
<th></th>
<th>$18^\circ$W–$10^\circ$W</th>
<th>$10^\circ$W–$0^\circ$</th>
<th>$0^\circ$–$10^\circ$E</th>
</tr>
</thead>
<tbody>
<tr>
<td>$m_{P_1}$</td>
<td>$-0.84$</td>
<td>$-0.49$</td>
<td>$-0.25$</td>
</tr>
<tr>
<td>$s_{P_1}$</td>
<td>$0.53$</td>
<td>$0.75$</td>
<td>$0.71$</td>
</tr>
</tbody>
</table>

$P_1$ (resp. $P_2$) which was markedly wet (resp. dry) on the two boxes. This is confirmed by the averages given in Table 2 showing identical values on the two boxes for $P_1$ and $P_2$, while $P_2$ is as dry (in fact drier) as $P_1$ on the western box but moderately dry on the central box.

The statistics of Table 2 also show that the correlation between the two boxes at the interannual scale is weak for all three periods. Interestingly enough the interannual correlation is as low for $P_1$ as for $P_3$, despite the fact that the means are identical on $P_1$ and very different on $P_3$. Thus, while there has been a clear decoupling of the Western Sahel and of the Central Sahel with respect to decadal trends since about 1990 – in opposition to the phasing of the preceding 20-year wet and dry periods – the two regions are displaying very little correlation at the interannual scale, whatever the period considered, whether wet or dry, and whether the two regions are in phase or not at the decadal scale.

In order to further evaluate how significant is this contrast between the two boxes the consistency of the interannual rainfall patterns within each box was tested by computing the rainfall indices at the $0.5^\circ \times 0.5^\circ$ scale. There are 100 $0.5^\circ \times 0.5^\circ$ boxes in a $5^\circ \times 5^\circ$ box and one can compare for each year the 100 $0.5^\circ \times 0.5^\circ$ box values to the average value obtained for the
The interannual fluctuations of the pressures of these two anticyclones being not systematically in phase is a strong factor of decorrelation between the Western Sahel rainfall and the Eastern Sahel rainfall.

Since there has not been observed a changing pattern of the average position and intensity of the two anticyclones over the past 20 years, the reason for the recent decoupling between the Western and the Central Sahel at the decadal scale must lie elsewhere. A possible factor, namely the role of decadal variations of the warming pattern of the tropical ocean, is highlighted in the recent paper of Hagos and Cook (2008). Regional climate simulations presented in this paper relate the 1980s Sahel drought to the warming of the Indian Ocean causing an anomalous anticyclonic circulation, the easterly branch of this circulation driving moisture away from the Sahel. The warming of the Tropical Atlantic enhances this effect by competing for the available moisture. The model sees some precipitation recovery in the 1990s linked to the warming of the Northern Tropical Atlantic, while the rest of the Tropical Atlantic cooled down. This induces a westward displacement of the divergence associated with a continuing warm Indian Ocean. The precipitation simulated by the model for the 1990s, despite some notable discrepancies with the observed rain fields, does produce a return to the rainfall of the control simulation [average SST of the 1950–1999 period] east to 0°.

Seasonal cycle

In their study comparing the seasonal cycle for the periods $P_1$ and $P_2$ on two transects located at 5°W and 2°E, Le Barbé et al., 2002 noted a shift in the timing of the continental rainy season setting up after the monsoon jump. The dry period, $P_2$, was characterised by an earlier occurrence of the seasonal rainfall maximum by about 20 days at all latitudes between 6°N and 10°N; this time shift progressively decreases when moving to the north; it is reduced to about one week at 14°N. In the following text, this variability of the seasonal cycle will be examined from two different perspectives: first by comparing the cycles of the two boxes for identical periods (Fig. 5) and secondly by comparing the cycles of the three periods over each box (Fig. 6). To that end, the mean seasonal cycles over each box and each period were computed by arithmetically averaging the individual seasonal cycles of the 11 stations of the western box, on the one hand, and of the nine stations of the central box, on the other hand.

As seen from Fig. 5, there are noticeable differences between the two boxes. The seasonal cycle of the 0°–5°E box for 1950–1969 may be considered as a reference for the Sahel, with five well identifiable stages: (i) a progressive increase of the 10-day rainfall from early April to the end of May, linked to the oceanic stage of the WAM (Lebel et al., 2003); (ii) a one month plateau; (iii) the monsoon jump during the last 10 days of June (Sultan and Janicot, 2003); (iv) the continental stage of the WAM characterised by a steady increase of the 10-day rainfall until it reaches its maximum of 70 mm at the end of August after a small break at the end of July; and (v) an abrupt retreat in one month, with residual rainfall in October. In the 15–10°W box, the seasonal cycle is fairly different,
with only three stages: a slow start until the end of May and then a steady and almost linear increase until early September when the 10-day rainfall peaks at a value close to 100 mm, followed by a fast retreat in September. The monsoon jump is remarkably absent.

The main features of these seasonal cycles are still visible on the right graph showing the seasonal cycles of the recent period, except on two points. In the 0°–5°E box the maximum 10-day rainfall is now observed during the first 10-days of August, instead of during the last 10-day period of August. Linked to this, the small break during the core of the rainy season is now observed at the end of July instead of beginning of August. In the 15°W–10°W box the 10-day rainfall maximum – diminished to 80 mm is now similarly observed during the last 10-days of August instead of during the first 10-days of September. Also, due to the persistence of low rainfall in the western box, the 10-day rainfall values are comparable over the two boxes until the beginning of August, when it peaks at 60 mm in the central box; this is in marked contrast with the wet period 1950–1969, when the western box 10-day rainfall was consistently larger by 15–20 mm than from mid-June to the end of August.

The 11-day moving average curves of Fig. 6 are providing a more detailed view of the mean seasonal cycles of each period. In the western box, the seasonal cycle of the most recent period is very close to that of the reference dry period P2, with relative deficits with respect to P1 over this period being 85 mm for P2 and 75 mm for P3. Thus three quarter of the 100 mm seasonal deficit remaining for the present period P3 is due to this earlier retreat of the rainy season. At the same time it must be noted that the overall length of the rainy season, defined as the period of daily rainfall larger than 1 mm/day is almost identical for the three periods (roughly from 10th May to 1st of October, that is about 110-days).

It is obvious from Fig. 6 that the rainfall deficit of periods P2 and P3, with respect to period P1 in the western box is more or less evenly distributed over all the rainy season after the 15th of June, while in the central box, the deficit is mostly concentrated after the 10th of August. This is confirmed by Table 3, giving the relative share of the total seasonal rainfall associated with the main stages of the rainy season. In the central box, 40% of the total seasonal rainfall (or 191 mm) is now falling between 23rd June and 10th August, against 35% (201 mm) in the 1950s–1960s, and 40% (191 mm) also after 10th August, against 46% (265 mm) in the 1950s–1960s.

These changes had a major impact on the hydrological regime of the Niger River (Fig. 7). Taking into account the 4–6 weeks delay between the local rainfall maximum and the first peak flow, one sees very well the effect of the stronger July rainfall of period P3 producing a September discharge similar to the one observed in the 1950s–1960s. Then the lower August rainfall for both P2 and P3 produces a diminution of about 200 m$^3$/s of the discharge with respect to period P1. The other interesting feature of this figure is the dramatic reduction of the second peak flow which is now 400 m$^3$/s lower than in the P1 years. Its timing has also advanced by 6 weeks. This second peak flow is the so-called “Guinean flood” produced by the July–September rainfall in western Mali and Guinea. Its decrease results from the strong August and September rainfall deficit over these regions where the rainfall recovery of

![Fig. 6. Comparison of the mean seasonal cycle over each of the two 5° × 5° boxes for the three periods.](image)

### Table 3

Proportion of rainfall falling during various stages of the rainy season for the different periods analysed here. The calculation is made for two different sequences of stages, one using the average onset date as a pivotal point of the seasonal cycle, the other based on the typical millet campaign calendar (120 days between 1st June and 30th September).

<table>
<thead>
<tr>
<th></th>
<th>1st April–22nd June (1st April–30th May)</th>
<th>23rd June–10th August (1st June–30th September)</th>
<th>11th August–31st October (1st October–31st October)</th>
</tr>
</thead>
<tbody>
<tr>
<td>$P_1$ Wet, western: 905 mm</td>
<td>0.10 (0.02)</td>
<td>0.33 (0.88)</td>
<td>0.56 (0.10)</td>
</tr>
<tr>
<td>$P_2$ Dry, western: 680 mm</td>
<td>0.11 (0.02)</td>
<td>0.37 (0.91)</td>
<td>0.51 (0.07)</td>
</tr>
<tr>
<td>$P_2$ Dry, western: 675 mm</td>
<td>0.10 (0.02)</td>
<td>0.34 (0.90)</td>
<td>0.55 (0.08)</td>
</tr>
<tr>
<td>$P_3$ Wet, central: 576 mm</td>
<td>0.18 (0.09)</td>
<td>0.35 (0.88)</td>
<td>0.46 (0.03)</td>
</tr>
<tr>
<td>$P_3$ Dry, central: 429 mm</td>
<td>0.20 (0.09)</td>
<td>0.38 (0.88)</td>
<td>0.41 (0.03)</td>
</tr>
<tr>
<td>$P_3$ Mild, central: 478 mm</td>
<td>0.19 (0.09)</td>
<td>0.40 (0.88)</td>
<td>0.40 (0.03)</td>
</tr>
</tbody>
</table>
the recent years is still weaker than over the Central Sahel (see values of $m_{\text{reg}}$ in Table 1).

Looking beyond the observational evidence of a clearer onset break in the central box it is worth noting that this was already noticeable in the paper of Sultan and Janicot (2003) looking in detail at the dynamics of the preonset and onset stages of the WAM. One dynamical explanation for this is the much sharper ocean-continent contrast existing at $0^\circ$–$5^\circ E$ and the smaller contrast in temperature and humidity at $10^\circ$–$15^\circ W$. The monsoon regime is thus better established and typical in the Central Sahel than in the Western Sahel. Furthermore the eastern edge of the Sahara heat low is positioned around or east of $0^\circ$ and the abrupt shift of the ITCZ at the time of the summer monsoon onset is associated with the heat low. It has been hypothesized that a reinforcement of the heat low in an area located west of $5^\circ E$ and south of the Atlas Mountains, increases the anticyclonic vorticity north of the Atlas Mountains, with a possible feedback acceleration of the seasonal cycle of the monsoon over West Africa. The fact that the shift of the ITCZ is much more abrupt in the Central Sahel than in the Western Sahel, means that the factors controlling this shift have a greater influence on the interannual rainfall variability there. It does not explain however why the seasonal rainfall maximum has durably shifted earlier in the summer. Further investigation on this subject is required.

Downscaling on the AMMA-CATCH-Niger meso-site

The AMMA–CATCH observing setup includes one mesoscale site densely instrumented in rainfall measurements since 1990 (Cappelaere et al., this issue). This meso-site (hereafter referred to as the ACN study area: $13^\circ$–$14^\circ N$, $1^\circ$–$3^\circ E$) is located right at the centre of the central box (Fig. 2). In the following we will use the data collected by the so-called climatological network of 30 recording rain gauges evenly distributed over the 16,000 km$^2$ meso-site in order to address the question of how similar/different the two areas are in term of interannual variability and seasonal cycle. This is tentative to evaluating how representative a meso-site of a dozen thousand km$^2$ is of a larger environment in term of rainfall regime. In a second step the higher time resolution data will be used to update our knowledge of the Central Sahel rainfall regime at smaller time steps.

Interannual variability of the ACN area compared to that of the central box

Fig. 8 compares the areal annual rainfall for the $5^\circ \times 5^\circ$ box and for the ACN area over the common period of observation (1990–2007). The areal rainfall was computed using the same best linear unbiased estimation algorithm (kriging) for the two areas (see Lebel and Amani, 1999 for details). The two series exhibit significant differences. First of all, the average rainfall over the $5^\circ \times 5^\circ$ box is smaller (479 mm) than in the ACN area (520 mm); however they have the same standard deviation of 73 mm.

Beyond this, one can notice significant discrepancies in the interannual fluctuations of the two series, with a coefficient of determination equal to 0.66. The largest difference between the two series is for 1998. That year the three Niamey stations of the ACN network recorded 1044 mm at Niamey1 (south of the town), 978 mm at Niamey2 (10 km to the north–west of the previous one) and 793 mm at Niamey3 (Airport, 15 km to the east of at Niamey2). The network used to monitor the $5^\circ \times 5^\circ$ box rainfall only includes the airport station, which means that the strong local
rainfall gradient recorded in the region of Niamey was not seen by this network. We are thus facing two cumulative effects explaining why the 1998 rainfall total estimates are so different over the two areas (672 mm over the ACN area against 537 mm in the $5^\circ \times 5^\circ$ box): (i) the exceptional maximum of the Niamey area was not sampled properly by the $5^\circ \times 5^\circ$ box network, leading to underestimating it; (ii) this area of rainfall maximum has a greater weight on the ACN rainfall than on the $5^\circ \times 5^\circ$ box rainfall since the ACN area has a smaller size. This illustrates how the strong spatial variability of the Sahelian rainfall analysed by Balme et al. (2006) may impact on the estimation of the rainy season. On the other hand, 1998 alone does not of course explain all the differences between the two series. When putting 1998 aside, the differences in averages is only slightly reduced (475 mm against 511 mm), and the coefficient of determination increases to 0.71, which leaves 30% of nonco-fluctuating variance between the two series. This nonco-fluctuation may come either from the spatial heterogeneity of the annual rain fields or from the location of the stations on the $5^\circ \times 5^\circ$ box. One can notice that six out of the nine stations are located in the north-eastern half of the box, where the average annual rainfall is smaller than in the south-west corner (Fig. 3). However, using a kriging algorithm to compute the areal rainfall minimises these sampling effects, so that most of the unexplained variance is probably attributable to the significant spatial variability of the annual rain fields. As a conclusion, the ACN area may be deemed to provide a fairly but far from perfect – representative vision of the annual rain fields. The calculation of meaningful and stable seasonal cycles would thus require to average over more than 20 years, but at the same time it was shown previously that there is a significant decadal variability of these seasonal cycles. It is thus a challenging task to define an average seasonal cycle for a given period and area in the Sahel.

Another interesting feature of Fig. 9 is the overall good phasing of the two signals, especially concerning the pause in June, the timing of the monsoon jump, the decrease in mid-July, the timing of the peak and of the retreat. Also worth noting are the pauses at 10-day intervals in the ACN curve in July and August, that could be related despite the averaging over 18 years – to the intra-seasonal modulation of convection described by Janicot and Sultan (2001).

**Point rainy days and point rain events**

Until now the rainfall regime was analysed in term of nonconditional statistics, that is zero rainfall realisations were included in the samples of rainy days. In the following, the intermittent nature of rainfall will be taken into account by separating nonrainy days – characterised by their rate of occurrence – and rainy days – characterised by their mean average rainfall.

**Seasonal cycles of occurrence rates and mean daily rainfall**

Fig. 10 is showing the rainfall regime of the three reference periods in term of conditional probability of rainfall larger than 2.5 mm. To obtain these curves, only the rainfalls larger than 2.5 mm are kept in each station series. This is equivalent to consider that a day is “significantly” rainy if at least one of the nine stations of the $0^\circ–5^\circ$E box records more than 2.5 mm that day.
The selected rainy days account for 86% of the total rainfall of the box for period P1; 81% for period P2 and 82% for period P3. These global statistics are an indication that wetter (drier) periods are characterised by a slightly larger (smaller) proportion of rain falling at daily rates larger than 2.5 mm. Computing the graphs of Fig. 10 with alternative thresholds of 0 mm and 1 mm produces very similar curves to those computed with the 2.5 mm threshold (not shown). This means that, while light daily rainfall smaller than 2.5 mm accounts for 15–20% of the total annual rainfall, fluctuations of their relative occurrence explains very little the decadal scale variability of the Central Sahel rainfall regime.

Fig. 10 clearly indicates that the major source of rainfall deficit is a lower occurrence of rainy days rather than a smaller intensity of daily rainfall, at least when comparing the wet period P1 to the two other periods. In particular the strong rainfall deficit of both periods P2 and P3 after the 10th of August is entirely caused by a dramatic decrease of the probability of rainy days, whereas the intensity of these rainy days remains similar for the three periods. A notable exception is the post onset period of July characterised by a dramatically lower mean daily rainfall for period P2 as compared to periods P1 and P3. This explains most of the total seasonal rainfall difference between period P2 and P3. Note also that the fluctuations of mean daily rainfall in June are not necessarily significant given the low probability of occurrence and thus the large sampling error associated to the calculation of the average daily rainfall. To summarize, Fig. 10 supports the conclusion that the first order signal explaining the wetter years 1950s–1960s is the much stronger probability of occurrence of rainy days, and this for the whole rainy season, with a particularly higher rate of occurrence between 10th August and 10th September. On the other hand most of the improvement of the rainy conditions over period P3 is linked to larger daily rain rates in July with an astonishingly similar seasonal cycle of occurrence for periods P2 and P3.

The various modifications described above are synthesised in Fig. 11 showing the statistical distributions of the conditional mean daily rainfall on the Central Sahel box for the two thresholds 0 mm and 2.5 mm.

At the 0 mm threshold, the distributions of the mean daily rainfall for the wet and the dry periods are identical up to the 99.8% quantile, while the recent period displays a tendency towards stronger daily rainfall beyond the quantile 80%: the quantile 98% is 58 mm for period P3 against 53 mm for periods P1 and P2. At the threshold 2.5 mm, the quantile 98% is 60 mm for period P3 against 54 mm for period P1 and 55 mm for period P2. A similar study on the Western Sahel showed that period P1 was characterised there by stronger daily rainfall. This is thus another difference in the modifications of the rainfall regimes of the two regions: the slight recovery of rainfall in the Central Sahel is mostly explained by stronger daily rainfall in the upper part of the distribution, while in the Western Sahel, both the number of rainy days and the mean daily rainfall remain in deficit. Overall however, when comparing periods P1 and P3, the main signal is the strong decrease of the number of rainy days everywhere in the Sahel after the monsoon jump; this signal is especially marked in August and during the first half of September over the Central Sahel.

From rainy days to rain events

Daily rainfall is the only source of information we have to analyse the fluctuations of the rainfall regime at the decadal scale. It is however a potentially biased representation of what is actually occurring in term of rainfall producing events. Because the organised Mesoscale Convective Systems (MCS’s) producing the largest share of rainfall in this region (Mathon et al., 2002) are long lasting and moving, the associated rain may fall either in one “meteorological” day (6:00–6:00) or in two days; on the other hand there is a significant probability of observing two distinct rain events in one day, especially for localised convective events or loosely organised MCSs. For many applications involving a realistic approach of the hydrological cycle it is much more relevant to work on rain events rather than rainy days, since the partition between infiltrated and runoff water is strongly conditioned by the amount of rain falling in a few hours at the event scale.

The ACN recording rain gauges, providing 5-min. rainfall series, will allow in the following to study the sampling effects related to the way one defines a rainy event (or a proxy to a rainy event, such as when using daily rainfall series), by comparing four series of data.

(1) The daily rainfall series obtained from the original 5-min. rainfall series at the 30 long-term ACN recording rain gauges are thresholded at 1.0 mm (i.e. only days recording rainfall larger than 1.0 mm at a given station are qualified as rainy days at that station). These daily rainfall series are considered as series of rain event proxies to simulate the situation when only daily rainfall data are available; they account for 95% of the total annual rainfall over the ACN area. Averaging over the 30 stations and over 1990–2007, monthly means are obtained (black circles in Fig. 12).
One can also define a rainy event at any given station from simple and classical criteria of duration and intensity. In our case, a rain total of 1 mm is the minimum required, and the rain event ends as soon as the rain stops for more than 30 min. These “point-events” also represent 95% of the rain recorded over the whole ACN region over the 18 years of observation (triangles in Fig. 12). When comparing the point-event curve to the daily curve, one can see that the daily rainfall is about 20% larger than the point-event rainfall and that, conversely, the occurrence rate of daily rainfall is 15–20% smaller than the occurrence rate of point events, which means that there are a significant number of small rains occurring more than once per day.

The availability of 30 recording rain gauges over the ACN area also allows for defining rainy events in a spatial perspective, in order to identify ground events that are as close as possible to the atmospheric events – i.e. the MCSs – producing most of the rainfall. To that end the criteria used to define “Mesoscale” rain events over the ACN area are adapted from those used to define a “point-event” as follows: (i) at least 30% of the rain gauges record more than 0.5 mm of rain; (ii) at least one station records 1 mm; (iii) when no rain is recorded at any of the network stations for more than 30 min, it is the end of the rain event. Since the study area covers 16,000 km², this definition means that, assuming some continuity, the rainy area covers at least 5000 km², a figure to be compared to the 5000 km² area of cloud cover below –40 °C used to define MCSs from METEOSAT imagery. Of course these two numbers are not directly comparable, but the systematic study of Mathon et al. (2002) showed that there is a globally good coherency between the ground mesoscale rain events identified from the criteria given above and the METEOSAT-defined organised MCSs: even though there is not a perfect one to one matching between these two sets of events, each population accounts for about 90% of the total annual rainfall, that is 5% – or 26 mm at the annual scale – less than the point events. The mean rainfall produced by these mesoscale events is consistently larger, by about 8%, than the conditional daily rainfall. This reflects the fact that a non-negligible proportion of rain fields associated with these mesoscale events are split between two consecutive rainy days.

The last series of rain events proxies is obtained by fitting to the point daily rainfall series the Compound Poisson Process (CPP) model (white noise, exponential distribution) described in Le Barbé et al., 2002. Since daily rainfall series is the most widely available source of information on rainfall...
worldwide, there is a wealth of literature on how deriving the statistical parameters of the underlying rainfall event distribution from daily rainfall series (see Bodo1 et al., 1987, for a review on CPP models and parameter estimations). The model chosen here is relatively simple and robust and the aim of the comparison is to check whether the results in Le Barbé et al., 2002 are consistent. The main advantage of this model lies in its additive properties, which allows for computing directly the monthly values of the rainfall distribution from the values obtained at the daily scale. These values are given in Fig. 12 along with their standard deviation of estimation errors and are compared to the observed point value statistics. The mean monthly values of the observed point-event rainfall are all lying in the 1-std confidence interval of the model values, indicating that the CPP model is an acceptable representation of the point rainfall process in this region.

Three main conclusions can be drawn from the curves of Fig. 12. First it is obvious that mesoscale rainfall events are definitely less frequent and more intense than point rainfall events which mix the local convective events and the mesoscale rainfall events. Since spatially sparse daily rainfall series make it difficult to separate local convective events from mesoscale rainfall events, a long-term mesoscale monitoring site such as the ACN site, whose data are used in conjunction with satellite data, proves to be very useful in studying rainfall properties at that scale. Secondly the CPP model is performing reasonably well at the monthly scale, providing point rainfall event statistics that are much more realistic than those derived directly from the daily rainfall series used as series of rain event proxies. Third, daily rainfall series should not be used directly for small catchment hydrological studies in this region.

The seasonal cycle in term of rain events

Having validated the CPP model on the ACN data, its parameters were computed over the 5° × 5° box and for each of the three periods P1, P2 and P3. The mean monthly values of the two parameters (occurrence rate and intensity of the rain events) of the CPP model are plotted with their standard deviation of estimation errors in Fig. 13. What comes out from this figure is a clear distinction between the P1 curve and the P2 and P3 curves on the graph of occurrence rates. On the opposite on the graph of mean event rainfall, the three curves are interlaced, with the values estimated for the periods P2 and P3 belonging to the 1-std confidence interval of the P1 values (except for the P3 September value). Even though the CPP model used in this study has some shortcomings in term of robustness when estimating its parameters, its results on the three periods are globally coherent, highlighting the fact that the main change of the rainfall regime since the 1950s and 1960s is the dramatically smaller occurrence rate of rain events in August and September, whereas the monthly averages of their point intensity did not change much. One point to consider however is the possibility of convective events of smaller extension during the dry periods, as suggested by Bell and Lamb (2006). The hypothesis is that the point average event rainfall remained more or less constant over the three periods considered here but that the decrease of the occurrence rate is at least partly linked to a diminution of the spatial extension of these events, implying that the areal average event rainfall would also have decreased.

Interannual variability of the mesoscale rain events

Point process models can be used to infer the rainfall regime in term of point rain events, but, as illustrated in Fig. 12 for the CPP model, they are of little use for inferring whether the spatial properties of the mesoscale events might have changed since the wet period P1. The only direct access to the mesoscale rain fields is provided by the ACN data, limiting our exploration capacity to the most recent period P3. Relating the interannual variability of the total seasonal rainfall to the number of mesoscale events – as defined in the previous section – on the one hand, and to the mean event rainfall, on the other hand, Fig. 14 indicates that over the recent period the average arial rainfall produced by mesoscale events is totally uncorrelated with the total seasonal rainfall. A contrario, the number of MCS events displays a significant correlation (r² = 0.43) with the seasonal rainfall. It can also be seen in Fig. 14 that the average MCS event rainfall value for 1998 is 13.5 mm, compared to an average of 11.0 mm over the 18 years of observation. Recomputing the linear correlation after removing 1998 from the sample, the coefficient of determination climbs to 0.51 for the number of events and decreases to 0.0 for the mean event rainfall. It can thus be concluded that the number of MCS rain events explains about half of the interannual variance of the seasonal rainfall, while the fluctuations of the associated arial event rainfall are not correlated with the fluctuations of the annual

![Fig. 13. Seasonal cycles of the parameters of the CPP model fitted to the daily series of observation at the synoptic Niamey station.](image-url)
Summary of results and discussion

While the drought of the 1970s and 1980s was more or less uniformly felt over most of West Africa with an average annual rainfall deficit displaying variations of ±20% around 200 mm, the recent period 1990–2007 is characterised by a more complex pattern. The study on the Sahel presented here (with a focus on the area encompassed by latitudes 11°N and 16°N) shows that: (i) in the Eastern Sahel (roughly east to 10°E), wetter conditions are observed since the end of the 1990s with the averages over the past 18 years being equal to their 1950–1989 counterparts; (ii) in the Central Sahel, the drought is somewhat abated, with a 1990–2007 average larger by 10% than the 1970–1989 average, but still lower than the 1950–1989 average; and (iii) over western Sahel, dry conditions are still prevailing, the interannual average of the recent period being equal to 675 mm on the 15–10°W box analysed here, to be compared to 905 mm for the wet period and 680 mm for the dry period 1970–1989.

The analysis of the seasonal cycle on the 5° × 5° reference box of the Central Sahel highlights that most of the rainfall deficit of the past 4 decades with respect to the wet 1950–1969 period is linked to a 1/3 decrease of the number of rainy days from the 10th of August to the 10th of September. As a consequence, the rainfall peak at the end of August totally disappeared and this accounts for 75% of the rainfall deficit of the dry years. On the other hand the slight recovery observed over the recent period in the Central Sahel is linked to larger rainfall in July by comparison to the dry 1970–1989 period; the rainfall regime of August and September remains unchanged between these two periods. This larger July rainfall of recent years on the Central Sahel is associated with larger daily rainfall in the upper part of their statistical distribution. This would indicate that the underlying process is different from the one that caused the end of the wet period over the whole Sahel.

The data collected by the 30 recording rain gauge network over the AMMA-CATCH-Niger (ACN) study area allowed for a finer study of the interannual variability and of the seasonal cycle over the years 1990–2007. The two annual rainfall series are correlated at 0.66 ($r^2$ value) only and the two average seasonal cycles are correlated at 0.57 ($r^2$ value, seasonal trend removed). Thus while there is an overall good coherence between the two series of measurements a significant variability in each series is not captured by the other. This does not prevent the ACN area to represent correctly the main features of the Central Sahel rainfall regime (interannual variability and seasonal cycle) but it has to be kept in mind that, for hydrological application and water budget studies, there is a significant gap of information involved when downscaling from $5° × 5°$ to $1° × 1°$ (or when upscaling from $1° × 1°$ to $5° × 5°$) at the yearly and monthly scales. The ACN setup is demonstrating here its utility in documenting the 10 km scale variability in this region and its potential interest for computing water budgets at the mesoscale.

The other main interest of the ACN data set is to document the rainfall regime at sub-daily time scales, thus allowing for a statistical characterisation of the rain events associated with the mesoscale convective systems. The average rainfall of these mesoscale rain events (from =14 mm in June and September to 16.2 mm in August) was shown to be larger than the average rainfall of rainy days, indicating that the rain produced by these mesoscale events is often split over two consecutive rainy days. The mesoscale-event rainfall is also larger than the average rainfall of rain events defined at the station scale (from =10 mm in June and September to 12.4 mm in August). The 25–30% difference between the mesoscale-event averages and the point-event averages is a measure of the spatial intermittency of these mesoscale events. A great deal of the rainfall variability in the region is explained by the variability of occurrence of these mesoscale rain events. At the decadal scale, the main modification of the rainfall regime of the 1970–2007 period with respect to the previous two wet decades (1950–1969) is the strong decrease of the number of rain events in August and during the first half of September. At the annual scale, half of the interannual variability on the ACN study area is explained by the number of mesoscale convective events; at the opposite the fluctuations of the mean rainfall produced by these mesoscale events displays absolutely no correlation with the total annual rainfall.

The mesoscale rainfall network of the Niger site is the oldest of the three mesoscale networks deployed in the framework of the AMMA-CATCH observing system, which aims at sampling the eco-climatic gradient of the region. This opens the way to three main directions of investigation: (i) with soon 20 years of data available on the Niger site, a robust analysis of the small scale factors – both in space and time – determining the interannual variability of rainfall in this region is now possible; in particular the signal recorded on the mean daily rainfall in July for the recent period on the daily data from the operational networks does not show up on the ACN data when looking at the interannual variability and this requires further investigation; (ii) despite the unequal length of the rainfall series on the three sites, a comparison of these fine scale factors between the sites will provide insight into their linkage with the climatic gradient; (iii) analysing the data produced by the observing system itself, is only a first step towards a better understanding of how the atmosphere dynamics factors may account for the variability of the rainfall regime over the range of scales that matter for the water cycle and in term of impact for the populations (water resources, agriculture, water related diseases).
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