Rainfall climatology of the HAPEX-Sahel region during the years 1950–1990

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Abstract

In the Sahel, rainfall is the single most important factor conditioning the hydrology and the climate, but comprehensive statistical analyses of the rainfall climatology in the region are rare. Yet, even though in the Sahel rainfall data are scarce by the standards of the temperate countries, it is shown here that it is possible to obtain a reasonably good idea of what the rainfall has been over Sahelian Niger for the past 40 years, both in terms of interannual variability and spatial distribution.

To that aim a statistical model is used, which decomposes the space–time fluctuations of long-term rainfall averages into the fluctuations of the mean event rainfall on the one hand, and of the mean number of rainfall events over any period of accumulation, on the other hand. This model is first applied to the analysis of monthly rainfall data over the whole of Niger. It is shown that the lasting drought which has affected Niger for more than 20 years is associated with a decrease in the number of rainy events, rather than to a decrease of the mean event rainfall, and that this decrease is more pronounced for the core of the rainy season. Because these fluctuations are not homogeneous over Niger, a 5°×5° zone centred on the HAPEX-Sahel 1°×1° square is selected in order to characterise more accurately the rainfall climatology of the HAPEX-Sahel area between 1950 and 1990. In comparison with what it was between 1950 and 1970, the average length of the rainy season has not changed significantly during the dry period 1970–1990. Rather, it is the decrease of rainfall in July and August that explains most of the diminution of the total annual rainfall over this part of the Sahel since 1970. The average number of rainy events in August was reduced by about 30%, while the mean event rainfall remained roughly constant. Finally, the analysis of the daily rainfall series for Niamey (which constitutes the longest record available in Niger, starting in 1905) enables the comparison of four periods of 20 years between 1910 and 1990. The period 1970–1989 appears to be by far the longest and most severe dry spell of the past century. Almost 90% of the annual rainfall decrease over this period is explained by the decrease of the mean number of rainfall events during July and August, while both the length of the rainy season and the mean event rainfall remained stable.

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1. Introduction

There is a striking mismatch between the extreme vulnerability of life in the Sahel to frequent rainfall deficits, and our limited knowledge regarding the rainfall climatology of the area. Speculations on the links between the drought and changes in the atmospheric circulation have been numerous since the early work of Bryson (1973). It has often been supposed that dry years over West Africa are caused by a southward displacement of the intertropical convergence zone (ITCZ) position (e.g. Winstanley, 1975; Kraus, 1977). This should go with a reduction in the length of the rainy season, a fact opposed by Nicholson (1981). To paraphrase Gregory (1982) it could be stated that, during the 80s, opinion about the nature and the statistical significance of the Sahelian drought has fluctuated almost as much as the rainfall conditions themselves. As underlined by Janicot and Fontaine (1993), this may be partly attributed to the variety of mechanisms involved. Recent progress in an approach pioneered by Lamb (1978a,b) has emphasised the link between the sea surface temperature (SST) anomalies at a global scale and the interannual variability of the Sahelian rainfall (Folland et al., 1986). Since, these anomalies are themselves the result of complex interactions between various atmospheric and continental processes which are not yet fully understood, there is ample room for additional studies aiming at a better characterisation of the interannual variability of the Sahelian rainfall.

Today, evidence has been gained that the drought of the 70s and early 80s was not bound to end rapidly. In Niger, as shown below for the period 1970–1993 (and in Lebel et al., 1995, for the years 1990–1993) the drought has lasted continuously for 25 years (1969–1993). Thus, it is now possible to carry out meaningful statistical studies to characterise the space–time rainfall distribution during this dry spell as compared with that of the previous wetter period. This paper is devoted precisely to such a study. It limits itself to the HAPEX-Sahel location and setup and is based on a statistical model tuned to the characterisation of the space–time fluctuations of long-term rainfall averages. The model allows for a distinction to be made between variations in the number of storms and variations in the mean storm rainfall. While our results are valid for the Central Sahel only, it is believed that the model could be applied fruitfully to the whole of the Sahel provided the data needed are collected. Moreover, as recognised by Janicot and Fontaine (1993), there does exist an overall zonal homogeneity of the Sahelian rainfall climatology. It is consequently reasonable to suppose that the major conclusions of this study apply to most of the Sahel.

A preliminary analysis of the monthly rainfall data for the whole of Niger shows that the recent dry period is characterised by a decrease in the number of rainy events, while the mean storm rainfall varies little. This first result has encouraged us to proceed further by applying our model to the analysis of a larger data set covering a 5° x 4° zone centred on the H-S 1° x 1° square and to the comparison of the rainfall time distribution in Niamey for periods of 20 years between 1910 and 1990. The results obtained over the whole of Niger were confirmed, which led us to undertake a detailed investigation of the rainfall fluctuations as a function of the latitude and the period of the year.

2. The data

The longest daily rainfall series available in Niger are those from Niamey and Zinder, which both started in 1905. Other stations began to operate in the following 20 years but it was not until after the second world war that the rain gauge network became dense enough to provide a reasonably good coverage of rainfall for climatological studies (see the reports

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CV, coefficient of variation, R, rainfall in mm.
Niger is a country extending from 12°N to 24°N, well into the Sahara desert. Nevertheless most of the population live in the south where regular agriculture is possible. Even though there are no absolute and undisputed latitude limits for the Sahel, Niger may thus be divided into two regions of roughly the same area: one, essentially desert, north of the 17° parallel, and one, Sahelian, south of this limit. Naturally the pattern of the raingauge network reflects the concentration of human activities in the south. Among 35 daily rainfall stations which operated continuously, with no more than 10% of missing data at any given station, between 1950 and 1990, 34 are located south of the 17° parallel and 33 south of the 15° parallel (Table 1 and Fig. 5 to be discussed later). Two different analyses will consequently be carried out. The first concerns the whole of Niger, and is based on the network listed in Table 1. The second analysis aims at providing a meaningful climatology for HAPEX-Sahel. It is thus restricted to a 5° x 4° zone (0°-5° in longitude, 11°-15° in latitude) centred on the H-S 1° x 1° square, hereafter referred to as the CSA (Central Sahel area). In order to minimise the border effects and to work on a larger sample of stations, all the raingauges available on a region extending 2° westward and 1° eastward of the CSA (Fig. 1) have been used for this analysis. Nineteen synoptic and climatic stations operated from 1950 to 1990 over this zone that covers south-east Burkina, south-west Niger and northern Benin. As well as these 19 stations, 57 additional raingauges have been used for the period 1950-1989, and a further 39 for the period 1970-1989. Fig. 1 shows that the enrichment of the network for the period 1970-1989 is especially important over the Niger part of the CSA. Working on a strip extending beyond the borders of Niger, into Burkina, allows us to obtain a sufficiently large sample for the computation of averages along latitudes for the period 1950-1969.

3. The leak distribution as a model of the rainfall regimes in tropical Africa

3.1. General presentation of the model

The regional analysis of rainfall regimes requires modelling which is coherent both in space and time. This goal is best achieved when the parameters of the model have a physical content. For any given period of rainfall accumulation, the mapping of these parameters then leads to a meaningful characterisation of how the rainfall varies in space. Also, the evolution of these parameters with the time of rainfall accumulation bears a direct climatic signification.

The model proposed here to describe and analyse the rainfall regime of Sahelian Niger, known as the leak distribution, is rooted in the renewal theory (e.g. Cox, 1964). To our knowledge it first appeared in the literature as a special case of the compound Poisson processes, proposed by Einstein (1937). It was then applied to rainfall analysis by Ribstein (1983), Le Barbé and Lebel (1989) and Le Barbé et al. (1989). The short presentation given below is taken from Le Barbé and Lebel (1989).

Let $R$ be the storm point rainfall accumulation. $R$ is supposed to be exponentially distributed, conditionally on the rainfall being non-zero (all the statistical analysis in the following applies to non-zero rainfall). Such an assumption was applied to storm rainfalls in southern United States by Smith and Schreiber (1974), followed by several
other authors. The climate of Arizona has some common features with that of Niger. It is semi-arid and storms are easily separable. It was therefore expected that the assumption of an exponential distribution of the event rainfall should hold for Niger as well. Fig. 2, where an exponential distribution is fitted to a series of 515 event totals recorded in Niamey by a tipping bucket raingauge between 1946 and 1981 (there are numerous missing data in this series, which explains why only 515 events are available over 36 years, see Bouvier, 1986, for details) clearly supports this hypothesis.

The c.d.f. (cumulative distribution function) of \( R \) may thus be written:

\[
F(r) = 1 - e^{-r/s}
\]

(1)

and the p.d.f. (probability density function) is:

\[
f(r) = \frac{1}{s} e^{-r/s}
\]

(2)

where \( s \) is the mean and the standard deviation of the storm rainfall. For climatically homogeneous periods, \( s \) is assumed constant.

The number of storms, \( N \), over any period \( T \) is further assumed to be a Poisson distributed random variable, that is:

\[
P(N) = \frac{e^{-\lambda_T} \lambda_T^N}{N!}
\]

(3)

where \( \lambda_T \) is the mean number of storms over period \( T \). For a given number of storms, \( N \), the rainfall accumulation over period \( T \), denoted \( R_T \), is then Pearson III distributed, (see e.g., Brunet Moret, 1969, for a comprehensive study of this distribution) with p.d.f.:

\[
f(R_T/N=n; n>0) = \frac{e^{-\lambda_T} \lambda_T^n}{n!(n-1)!} \sum_{i=0}^{n-1} \frac{(\lambda_T U_T)^i}{i!}
\]

(4)

where \( U_T = \frac{R_T}{s} \).

Defining \( U_T \), as the scaled variable \( R_T/s \), it becomes:

\[
f(U_T/N=n) = \frac{e^{-\lambda_T} \lambda_T^n}{n!(n-1)!} \sum_{i=0}^{n-1} \frac{(\lambda_T U_T)^i}{i!}
\]

(5)

The marginal distribution of \( U_T \), letting \( n \) vary from 1 to infinity is then:

\[
f(U_T) = \sum_{n=1}^{\infty} \frac{e^{-\lambda_T} \lambda_T^n}{n!(n-1)!} \sum_{i=0}^{n-1} \frac{(\lambda_T U_T)^i}{i!}
\]

(6)

or:

\[
f(U_T) = e^{-\lambda_T} \lambda_T U_T \sum_{n=0}^{\infty} \frac{(\lambda_T U_T)^n}{n!(n+1)!}
\]

(7)

Let \( J = \sum_{n=0}^{\infty} \frac{(\lambda_T U_T)^n}{n!(n+1)!} \), then \( J \) may be written:

\[
J = I_1(2\sqrt{\lambda_T U_T}) / \sqrt{\lambda_T U_T}
\]

(8)

where \( I_1(2\sqrt{\lambda_T U_T}) \) is the modified first order Bessel function. The distribution of \( R_T \) is thus given by:

\[
f(U_T) = e^{-\lambda_T} \lambda_T U_T \frac{I_1(2\sqrt{\lambda_T U_T})}{\sqrt{\lambda_T U_T}}
\]

(9)

for \( U_T > 0 \)

with \( U_T = R_T/s \), and:

\[
F(0) = e^{-\lambda_T}
\]

(10)

The first three centred moments of the leak distribution are:

\[
K_1 = E[R_T] = \lambda_T s
\]

(11)

\[
K_2 = E[(R_T - E[R_T])^2] = 2\lambda_T s^2
\]

(12)

\[
K_3 = E[(R_T - E[R_T])^3] = 6\lambda_T s^3
\]

(13)

which allows computation of the coefficient of variation and the skewness coefficient:

\[
CV = \frac{\sqrt{3}}{2} \sqrt{\lambda_T}
\]

(14)

\[
\gamma_1 = 3 \sqrt{2} \left( \frac{2\lambda_T}{\lambda_T} \right)^{1/2}
\]

(15)

Since the two parameters of the leak distribution are the mean event rain depth and the mean number of rainy events over the period \( T \) considered, their mapping permits a study of how the variation in the average rainfall is related to the fluctuations of these two factors. In this respect our model is comparable to the one used by Garbutt et al. (1981). However, and as far as the hypothesis of the time stationarity of the mean event rain depth holds, it has the additional advantage of making possible the deduction of the distribution of rainfall over any period \( T' = kT \) from the distribution computed for the period \( T \) (\( s' = s \) and \( \lambda' = k\lambda_T \)).

3.2. Inference of the parameters

The leak distribution (hereafter denoted LD) can be fitted to a series of rainfall cumulated over any given period \( T \), provided that: (i) \( T \) is long as compared with the mean duration of a rainy event; (ii) the rainfall process is time-stationary over \( T \) (that is \( s \) is constant). This latter assumption may be somewhat relaxed under certain conditions, but we will not make use of this possibility here. The LD fitting is based on the inference of two parameters, LD1 and LD2, that are used as estimates of \( s \) and \( \lambda_T \), respectively. Although this inference, using classical approaches such as the moment or maximum likelihood (ML) methods is relatively easy to carry out computationally speaking (see Babusiaux, 1969, and Ribstein, 1983, for the formulation of these algorithms), it may involve some distortion in the description of the underlying rainfall process. The major cause for such distortions is the correlation existing between the two estimates for both methods cited above (Babusiaux, 1969):

\[
r(LD1_T, LD2_T) = \frac{2\lambda_T + 1}{2(\lambda_T + 1)}
\]

(16)
The correlation coefficient between LD1T and LD2T depends on AT only, and it tends towards 1 for large values of λT, while it tends towards 0 for small values of λT. Thus, λT should be kept as small as possible, by choosing short periods for T. In practice two durations are most often used in rainfall climatology studies, that is 10 days and 1 month, for reasons linked either to modelling purposes or convenience (data availability). For a monthly mean number of events of 9 (an average value for September in the Sahel), the r square is equal to 0.90 when working on monthly totals and 0.77 when working on 10-day totals. For a monthly mean number of events of 6 (an average value for June in the Sahel), the r square is equal to 0.96 when working on monthly totals and 0.77 when working on 10-day totals (a comprehensive study of the correlation effect and associated estimation error variances of LD1T and LD2T will appear in a forthcoming paper). Since the underlying hypothesis in the building of the LD is that s and λT are independent this correlation is embarrassing. In particular, when comparing hvo stations or two different periods of observation at the same station in order to detect possible modifications in the rainfall climatology, the interpretation of changes in the value of s and λT will be made more difficult due to possible compensation effects in the estimation procedure (thus exaggerating or attenuating the real increase, or decrease, of one parameter).

Therefore an alternative method is proposed utilising the average number of dry days d0 over a period T of d days, which is information that is not used in the moment and ML methods. Eq. (11) gives direct access to an estimate of λT as:

\[ LD2T = -LN \left( \frac{N_0}{N} \right) \]  

where N_0 is the number of observed dry sequences of length T and N is the total observed number of sequences of length T in the sample. The computation of LD1T is then based on a no-bias condition:

\[ LD1_T = \frac{\bar{R}_T}{LD2} \]

where \( \bar{R}_T \) is the sample mean, used as an estimate of E[R_T^2] (19)

The ratio N_0/N is an estimate of the probability, denoted F0, of zero rainfall over the period T. The obvious advantage of this F0 algorithm is that LD2_T is estimated independently of LD1_T and that it makes use of important information on the Poisson process. The compensation effects mentioned above for the moment and ML methods are thus reduced, as demonstrated by comparing the correlation coefficient of the estimates, given by the following formula:

\[ r(LD1_T, LD2_T) = \left(1 + \frac{\lambda T}{e^{\lambda T} - 1} \right) \]

with the correlation of the ML and moment estimates given in Eq. (17).

For small values of \( \lambda_T \) the correlation coefficient tends towards zero. It also tends towards 1 when \( \lambda_T \) tends towards infinity. There is thus a major advantage to working on samples characterised by a low value of \( \lambda_T \). Selecting \( T = 1 \) day is an efficient way in that direction. In the Sahel \( \lambda_T (T = 1 \) day) will remain below 0.6 and the correlation coefficient between the estimates of the daily LD1 (denoted LD1) and LD2 (denoted LD2) will be kept below 0.3 in absolute value (r(LD1, LD2) = -0.27 for \( \lambda_T = 0.6 \) and r(LD1, LD2) = -0.14 for \( \lambda_T = 0.3 \)). Consequently the following procedure for estimating LD1_T and LD2_T has been followed here:

1. estimation of the daily values of LD1(i) and LD2(i) for each day i of the period T of interest, estimating F0 as:

\[ F0 = \frac{d_0}{d} \]

where d_0 is the average number of dry days during period T of d days;

2. estimation of LD1_T and LD2_T as:

\[ LD1_T = \frac{1}{d} \sum_{i=1}^{d} LD1(i), \text{ and} \]

\[ LD2_T = \frac{1}{d} \sum_{i=1}^{d} LD2(i) \]

In Section 4 below the period T considered is 1 month (\( d = 30 \) or \( d = 31 \)). When working on a sample of 20 years, F0 is thus determined from a total number of 6000 daily observations. In Sections 5 and 6, a moving window of 11 days is considered and there are still 2200 observations to estimate F0 from a sample of 20 years. Consequently, the two main advantages of the method are (i) ensuring a low correlation between the parameter estimates and (ii) allowing a robust estimation of F0 and thus of LD2_T. Its weakness lies in using daily rainfall data for the determination of the number of dry days, since a rainfall event may be spread over two different days, even if lasting for a few hours only. Nonetheless the probability of such an occurrence is reduced by the fact that daily readings are made at 8:00 a.m., an hour of lowest rainfall probability due to the diurnal cycle of convection. Furthermore a study carried out on the EPSAT-Niger data has shown that the average duration of a convective storm in the Sahel is around 5 h (Lebel et al., 1997, this issue).

It should be noted that, since the sum of n Poisson distributions is still a Poisson distribution the LD model is valid for all durations between 1 day and 1 month. However, since the LD2 (\( \lambda T \)) estimate is not a linear function of the data, the LD2_T estimate is generally not identical whether computed directly from monthly rainfall data or as the sum of the 30 (or 31) daily LD2 estimates. This is the price to pay for obtaining an effective parameter directly from monthly data which are easier to access and to process. Comparisons made for a few stations belonging to both the samples used in Sections 4 and 5 have shown that differences do not exceed 10%, an accuracy which is sufficient for a global analysis of monthly rainfall regimes, as carried out in Section 4.


As for the whole of continental West Africa, the rainfall regime of Niger is under the dependence of the north–south migration of the ITCZ. North of the ITCZ, high pressures originating from the Sahara prevent any rainfall, except in the case of rare descents of cold
air from the north during the boreal winter. Rain thus occurs over a given area only after the ITCZ has moved past this area towards the north. The maximum northward extent of the ITCZ is reached in August, corresponding to the period of maximum rainfall over the Sahel. The movement of the ITCZ northern edge is not regular which is the cause of often erratic starts of the rainy season. Even when the rainy season is well established sudden retreats southward of the ITCZ are not uncommon. The annual rainfall depends heavily on the duration of the rainy season, that is on the mean time during which the ITCZ is positioned north of the point considered. As a consequence, the yearly average isohyets are oriented east–west (Fig. 3), with a north–south gradient of roughly 1 mm km⁻¹ (the average yearly rainfall varies from 140 mm in Agadez to 720 mm in Gaya, separated by 560 km in latitude). Note, however, that in the eastern part of the country (region of Zinder and Lake Chad) the isohyets dive south-eastward.

The 40 years average over the H-S study area ranges from about 600 mm in the south (13°N) to a little less than 500 mm in the north (14°N). The average of Niamey over this period (564 mm) is close to the areal average which can be computed by integrating the isohyetal map of Fig. 3 over the H-S square (560 mm).

4.1. The Joseph effect affecting the rainfall in Niger

The last 40 years have witnessed a modern remake of the 7 dry–7 wet years sequence that introduced Joseph as the first climatologist of Africa during the pharaonic era. In their paper "Nest, Joseph and operational hydrology", Mandelbrot and Wallis (1968) have coined the expression "Joseph effect" to "designate the fact that a long period of high or low precipitation can be extremely long indeed". The lasting drought which has plagued West Africa since the end of the 60s was documented early in the 80s (Nicholson, 1980, 1981; Lamb, 1982, 1983). This drought continued throughout the 80s and it constitutes the most recent example of this Joseph effect, which Mandelbrot and Wallis believed was an intrinsic characteristic of rainfall regimes, rather than an unexpected accident. This drought has been especially severe over Niger, as may be seen from Fig. 4, where the fluctuations of the following scaled rainfall index \( I_R(k) \) are shown:

\[
I_R(k) = \sqrt{\frac{1}{T} \sum_{i=1}^{T} \left[ R(i, k) - \bar{R}(i) \right]^2}
\]

where \( R(i, k) \) is the annual rainfall at station \( i \) for year \( k \), and \( \bar{R}(i) \) is the annual average over the reference period for station \( i \). \( T \) is the total number of stations used for the computation.
(a subset of 20 stations homogeneously spread over the territory was chosen, out of the 35 stations of Table 1, to build the graph of Fig. 4).

As previously noted by Hubert and Carbonnel (1987) and Hubert et al. (1989) the persistence within each period is striking: all years before 1968 have a positive rainfall index, while all years, but one, after 1968 have a negative rainfall index. Fig. 5 shows that this general decline of the rainfall over Niger is equivalent to a 150-km southward shift of the annual isohyets during the dry 1968–1989 period. However this general trend masks important differences from one station to another. Excluding the two northern-most stations (Guba and Agadez, north of 16°N) the relative rainfall decrease between the period 1950–1967 and the period 1968–1989 varies from 7% (Gaya) to 43% (Goure, Tanout) (Table 1). The maximum absolute decrease reaches 220 mm in Filingue (from 453 mm to 232 mm) but it was only 60 mm in Gaya (from 553 mm to 793 mm).

A better understanding of this phenomenon can be obtained only by analysing how it has affected the rainfall distribution within the rainy season: has the contribution of each month remained stable or has it changed? Secondly, was the proportion of strong versus weak rains the same? Finally, was the same proportion of strong versus weak rains found in the northern part of the country as compared with the southern part? Fig. 5 shows that on average there is a 100-mm southward shift during the 1970–1989 dry period.

Fig. 5. Comparison of the annual isohyets between the period 1950–1969 and the period 1970–1989. On average there is a 100-mm southward shift during the 1970–1989 dry period.
medium or weak rainfall events modified? We seek to answer these questions by examining the evolution of the parameters of the leak distribution fitted to monthly totals, that is, we assume in a first step that the rainfall process is time stationary over a period of 1 month. Although calendar months are not necessarily meaningful from a climatic point of view, they constitute the most convenient entities to reach a synthetic characterisation of the changes into the intra-seasonal rainfall distribution. An alternative to the monthly average approach will be proposed in Section 5 to analyse with more detail the space-time fluctuations of rainfall over Central Sahel.

4.2. Modification of the monthly rainfall during the recent dry period

The drought in Niger started in 1968, as can be seen from Fig. 4. Lebel et al. (1995, 1997) have shown that it lasted until 1993. A break was observed in 1994. In other parts of the Sahel, especially those forming the western part of the Central Sahel area (CSA) studied below, the start of the drought was delayed by 1 year or 2. By the beginning of the 70s though, the drought was well established everywhere, as shown by Gregory (1982). In order to deal with periods of equal duration when comparing the wet and dry spells that occur in succession, we will thus consider in the following a wet period (J1) of 20 years between 1950 and 1969, and a dry period (J2) of 20 years between 1970 and 1989 (as seen in Fig. 4, the years 1968 and 1969 were moderately dry in Niger anyway). Ending the period of analysis in 1989 will provide an independent reference to be used by Lebel et al. (1997) for the analysis of the rainfall climatology during HAPEX-Sahel (1990-1993).

\begin{figure}[h]
\centering
\includegraphics[width=\textwidth]{Ayorou_June_1970-1989}
\caption{Ayorou - June 1970-1989}
\end{figure}

\begin{figure}[h]
\centering
\includegraphics[width=\textwidth]{Figures}
\caption{Fig. 6. Leakk, Pearson III and Gaussian distributions fitted to the June monthly rainfall series of Ayorou (1970-1989).}
\end{figure}

The leak distribution has been fitted month by month for the 35 stations of Table 1, separately for the two periods J1 and J2, using the FO algorithm. Six months, May through October, were selected, which account for 99% of the annual rainfall (Table 2). For May and October the number of zero values is too large to obtain reliable fits over periods of 20 years only. The inference of the leak parameters was thus restrained to the 4 most rainy months (June, July, August, September) which still account for more than 90% of the annual total in the south and more than 95% in the north of the country.

The fit of the leak distribution was then compared with the fits of the Pearson III (two parameters) and Gaussian distributions, which means a total of 280 samples (35 stations, 4 months, 7 periods). Since the small size of the samples does not allow for a \( \chi^2 \) testing, an alternative test proposed by Brunet Moret (1978) was used for this comparison. For more than 80% of the samples the better fit was obtained with the leak distribution, while the Pearson III

\begin{figure}[h]
\centering
\includegraphics[width=\textwidth]{Figures}
\caption{Fig. 7. Isoline maps of the leak distribution parameters for the periods 1950-1969 and 1970-1989. June monthly rainfall, LD parameters estimated by the number of dry days algorithm.}
\end{figure}
distribution had a better fit on 15% of the samples. An example of the behaviour of each distribution is given in Fig. 6 for the station of Ayorou. The leak distribution has a greater ability to reproduce correctly the probability of zero rainfall, and its asymmetry is intermediate between that of the Gaussian and that of the Pearson III distributions. In fact, for these three distributions, the ratio \( \gamma_1/\text{CV} \) is independent of their parameters: \( \gamma_1/\text{CV} = 0 \) for the Gaussian distribution, \( \gamma_1/\text{CV} = 3/2 \) for the leak distribution, \( \gamma_1/\text{CV} = 2 \) for the Pearson III distribution. The overall better fits of the leak distribution are an empirical confirmation of the suitability of this model for characterising the rainfall regimes in the Sahel.

The spatial variations of the two LD parameters are shown in Fig. 7 and Fig. 8 for the months of June and August, respectively. In June both LD1.30 (the mean event rainfall) and LD2.30 (the mean number of events) decrease when shifting from period J1 to period J2. However, the decrease of LD2.30 is remarkably well organised in space, the isolines shifting southward in a way that is very comparable to the shift of the yearly rainfall shown in Fig. 5. The changes in the LD1 map from J1 to J2 are more erratic, and smaller. In August the general pattern is similar. The LD2.30 isolines are oriented east--west and, for a given latitude, their value diminishes by 3 to 5 units between the J1 and J2 periods. In East Niger the LD1.30 value is also reduced, but by a lesser proportion than LD2.30. Elsewhere the variations of LD1.30 are negligible as compared with those of LD2.30. It is unclear whether this different behaviour is due to sampling effects when computing the LD parameters, or to regional differences in the consequences of the drought on the rainfall distribution. Nevertheless Fig. 8 supports globally the idea that much of the August rainfall decrease is attributable to an average decrease of 20% in the mean number of events. For half of the stations this decrease is even larger than 25%.

Thus, at the monthly scale the link between the drought and the decrease in the number of storms emerges. We will now focus on the CSA defined in Fig. 1 to analyse the changes of rainfall climatology with finer resolution. The time resolution will be improved by considering the \( \lambda \) parameter of the Poisson process as a continuous function of time, and the spatial resolution will increase as well, thanks to the large number of stations available in Burkina. This will also allow us to work on a climatically more homogeneous region by excluding East Niger from the data set.


5.1. Conditions of the analysis

To analyse the rainfall climatology over the CSA in more detail, two modifications to the approach taken in Section 4 are proposed.

First, the stations used for each period are different, in order to take advantage of the improvement of the network over Niger during the end of the 60s (Fig. 1).

Another important modification is related to the computation of the parameters of the leak distribution. In Section 4, the rainfall process was assumed to be approximately time stationary at the monthly scale, thus allowing the direct inference of the average monthly LD parameters. However, the relatively short duration of the rainy season, especially in the north of the CSA, implies that the rainfall in this region is strongly dependent on the ITCZ migration and that the hypothesis of the monthly stationarity is somewhat unrealistic when a more detailed analysis is undertaken. Therefore, it was decided to work on 11-day moving averages in order to account for the time evolution of the parameter \( \lambda \) of the Poisson process. In this method the statistics are computed over a 11-day window that is moved day by day. The Poisson process is considered stationary over any 11-day period, allowing for the computation of the two LD parameters of that period, using the inference algorithm based on the estimation of the number of dry days \( F_0 \). These two parameters are used as estimates of the central day (the sixth day of the period) parameters. A pair of LD parameters (LD1, LD2) is thus computed for each day, based on the information of the surrounding 10 days, producing a daily representation of the LD parameter fluctuations over the rainy season.
Fig. 9. Isoline maps of the seasonal (1st May through 31st October) rainfall parameters for the periods 1950–1969 and 1970–1989.

Fig. 10. Time fluctuation of the event rainfall and mean number of events during the rainy season (three stations of Niger over the periods 1950–1969 and 1970–1989).
5.2. The annual rainfall before and after 1970

In Fig. 5 the rainfall deficit of the years 1970–1989, in comparison with the 1950–1969 rainfall, amounts to about 100 mm in Northern Niger, increasing southward to almost 200 mm in the region of Niamey. Similar numbers are found when comparing these two periods over the CSA (Fig. 9). This confirms the extent and generality of the drought that set in at the end of the 60s. The coefficient of variation map shows that this sharp decrease of the average has been accompanied by an increase of the relative year to year variability, a fact that is not without consequences for vegetation production. The annual number of rainfall events has also been computed for each station using the following procedure. The LD distribution is fitted to the monthly rainfall distribution by the FO

algorithm for each of the 6 months, May to October. Summing, the monthly number of rainfall events (LD2) parameter yields the seasonal number of rainfall events which is mapped in Fig. 9.

The close link between the Sahelian rainfall decrease and the diminution of the number of rainfall events is confirmed. On average, the seasonal number of rainfall events has decreased by 12 to 15 units year \(^{-1}\), while the relative decrease of both the seasonal rainfall and the mean seasonal number of events is about 25%.

5.3. Space-time fluctuations within the rainy season

Six synoptic stations were selected to characterise the latitudinal rainfall gradient. Those stations are: Kandi (11.14°N; Benin), Gaya (11.88°N; Niger), Diapaga (12.06°N; Burkina), Niamey (13.48°N; Niger), Tillabery (14.20°N; Niger), Ayorou (14.75°N; Niger). The moving window method has been applied to compute the daily LD parameters for the six stations. The results of the computation are summarised in the time charts of Fig. 10 for Gaya, Niamey and Ayorou. In the south (Gaya) the mean number of events (LD2) does not change much between the periods J1 and J2. Rather, it is the mean event rainfall (LD1) that seems to have decreased. At Kandi (not shown in Fig. 10), neither LD1 nor LD2 present any systematic change between the periods J1 and J2. For all the other stations the LD2 parameter is the one that changes most between the two periods. The decrease of the number of events is especially important for the core of the rainy season in Diapaga and Niamey, while it affects the whole season in the North (Ayorou in Fig. 11). Note that, except in Ayorou, the rainy season does not appear to have been shortened during the dry years, an observation already made by Nicholson (1981). Regarding the mean event rainfall, no obvious pattern comes out. With the notable exception of Gaya the fluctuations are erratic, they differ from one station to another and along the year. A sharp decrease of the mean event rainfall is observed for Niamey in September and October. A careful analysis of data from other stations is needed to explain this phenomenon which is observed neither at Diapaga, nor at Tillabery and Ayorou. A possible numerical effect cannot be excluded at this stage.

The variations of the LD2 parameter are of a significant magnitude if compared with the confidence interval of the estimates. For the FO algorithm the variances of the estimates are given by the following expressions:

\[
\text{Var}(\text{LD}_1) = \left( \frac{1}{\lambda_f} \right)^2 \frac{e^{\lambda_f} - 1}{N}, \quad \text{and} \\
\text{Var}(\text{LD}_2) = \frac{e^{\lambda_f} - 1}{N}
\]

where \(N\) is the sample size.

The coefficient of variation is thus the same for the two estimates:

\[
\text{CV}(\text{LD}_1) = \text{CV}(\text{LD}_2) = \left( \frac{1}{\lambda_f} \right) \sqrt{\frac{e^{\lambda_f} - 1}{N}}
\]
and the relative 90% confidence interval (CI90) may be computed as 1.64 CV, or:

\[
LD_1 \pm 0.82 \left( \frac{LD_1}{LD_2} \right) \sqrt{\frac{1}{N}}, \quad \text{and}
\]

\[
LD_2 \pm 0.82 \sqrt{\frac{1}{N}} \quad (28)
\]

since LD_1 and LD_2 are used as estimates of \( s_f \) and \( \lambda_f \), respectively.

For \( T = 1 \) day, \( \lambda_f \) and its estimate LD2 are kept below 1.0; when LD2 = 0.5, CI90 (LD2) = 0.6, and CI90 (LD1) = 0.1, 0.17, and CI90 (LD1) = LD1 \pm 0.28LD1. While in Fig. 10 the variations of the LD1 parameter are within the limits of the 90% confidence interval, it is clear that the 1970–1989 values of LD2 are below the lower bound of the 1950–1969 CI90 for Niamey (LD2(1950–1969) = 0.6) and Ayorou (LD2(1950–1969) = 0.5) in August. A similar result was obtained for the three other stations (not shown). For Gaya it is quite different, the relative CI90 (LD2) = 0.17, and CI90 (LD1) = 0.28LD1. For LD2 it is quite different, the relative fluctuations of LD1 being more important than those of LD2. There is no obvious explanation of this singularity.

Since the six time graphs obtained for each station are spread more or less regularly from south to north, it has been possible to interpolate them spatially in order to produce space-time maps for both parameters LD1 and LD2 and both periods J1 and J2. These maps are obtained for the three other stations (not shown). For Gaya it is quite different, the relative CI90 (LD1) = 0.17, and CI90 (LD2) = 0.28LD1. While in Fig. 10 the variations of the LD1 parameter are within the limits of the 90% confidence interval, it is clear that the 1970–1989 values of LD2 are below the lower bound of the 1950–1969 CI90 for Niamey (LD2(1950–1969) = 0.6) and Ayorou (LD2(1950–1969) = 0.5) in August. A similar result was obtained for the three other stations (not shown). For Gaya it is quite different, the relative CI90 (LD2) = 0.17, and CI90 (LD1) = 0.28LD1. For LD2 it is quite different, the relative fluctuations of LD1 being more important than those of LD2. There is no obvious explanation of this singularity.

Since the six time graphs obtained for each station are spread more or less regularly from south to north, it has been possible to interpolate them spatially in order to produce space-time maps for both parameters LD1 and LD2 and both periods J1 and J2. These maps are given in Fig. 11. The critical disappearance of the dome characterising the LD2-J1 map in August is the most striking feature of this figure. Below 12°N the variations of LD2 are relatively small, while in the north they affect the entire period from June through mid-September. By comparison, changes in the mean event rainfall are small, with the range of 11 to 13 mm per event covering the majority of the map for both the J1 and J2 periods.

![31 DAY CUMULATIVE RAINFALL (mm) 1950 - 1969](image1.png)

![31 DAY CUMULATIVE RAINFALL (mm) 1970 - 1989](image2.png)


The combination of the maps in Fig. 11 and the integration over moving monthly periods permits us to obtain space–time maps of monthly rainfall (Fig. 12), synthesizing the change of rainfall regime which occurred over Central Sahel over a very short time span (1 or 2 years) at the end of the 60s. The three main facts noted throughout this section are clearly apparent: little variation during the setting in of the rainy season; dramatic decrease of the rainfall during the core of the rainy season, with a reduction by two thirds in August, the space–time pattern of the distribution being preserved; almost no change at the end of the rainy season.


The study of the Niamey rainfall series allows comparison of the rainfall climatology of the past 40 years with that of a longer period. It appears that, despite the particularity of the ‘Joseph sequence’ between 1950 and 1990, the main statistics of the series 1905–1989 and 1950–1989 are similar (Table 3). Behind this similarity of the global statistics, however, are hidden some important differences, which become apparent when the period 1910–1989 is divided into four subperiods of 20 years each. Using the 11-day moving window approach, the LD parameters have been computed at a daily step for each subperiod (Fig. 13 and Fig. 14). The most striking feature of these two figures is the great variability of the LD2 curves (daily mean number of events) as compared with that of the LD1 curves (mean event rainfall). The value of LD1 remains between 10 and 13 mm during the core of the rainy season (July through mid-September, or Days of year 182 to 258), for all four subperiods considered, except the first one. Intra-seasonal fluctuations are smooth. On the contrary, both the peak and average of the LD2 parameter change markedly from one subperiod to another. The peak is slightly above 0.5 for the period 1910–1929. It increases to 0.6 during the years 1930–1949 and to 0.7 in 1950–1969, before diving to 0.5 during the current dry spell (1970–1989). The intra-seasonal fluctuation is different from that of the LD1 parameter. The period of maximum is of 1 month only (roughly corresponding to the month of August). The August average is slightly above 0.5 for 1910–1929, around 0.55 for 1930–1949, around 0.6 for 1950–1969, well below 0.5 (0.45) for 1970–1989. This maximum is reached progressively and almost linearly, starting from a low value of 0.1 at the beginning of May (which means the equivalent of one event in 10 days). The post-maximum decrease is sudden and rapid. These variations are similar to the pattern of daily rainfall itself, which is known to increase slowly during the onset of the rainy season.

<table>
<thead>
<tr>
<th>Year</th>
<th>Mean</th>
<th>Standard deviation</th>
<th>Coeff. var. (CV)</th>
<th>Coeff. asym. (( \gamma_1 ))</th>
<th>CI90 (LD1)</th>
<th>CI90 (LD2)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1905–1989</td>
<td>502</td>
<td>1.35</td>
<td>0.24</td>
<td>0.39</td>
<td>1.6</td>
<td>11.3</td>
</tr>
<tr>
<td>1950–1989</td>
<td>564</td>
<td>1.22</td>
<td>0.26</td>
<td>1.3</td>
<td>10.9</td>
<td>52.2</td>
</tr>
</tbody>
</table>
By contrast the mean event rainfall increases rapidly between the beginning of May and June and then remains somewhat stable for almost 3 months.

Obviously, most of the rainfall fluctuations observed from one subperiod to another and within the rainy season are related to those of the mean number of events rather than to the variability of the mean event rainfall. Apart from the strong decrease of the mean number of events, one thing that may have changed during the dry spell is an earlier retreat of the rainy season: the decrease of the LD1 parameter happens 15 days earlier than during the years 1950–1969. Of course there remain sampling and numerical effects that should make one cautious when interpreting such curves. For instance, the LD1 peak of the 1910–1929 curve at the end of July, lasting for about 20 days, compensates for a hole in the LD2 curve. It is tempting to correct the two curves to make them more coherent with those of the other periods, but we have no objective indication to guide us in doing so. Even if such a correction is made, the mean event rainfall would remain above that of the following periods. It may well be that the rainfall regime during the first quarter of the century had some significant differences from the one that prevailed later, characterised by an overall stability of the mean event rainfall during the core of the rainy season.

Consider that the July–August period accounts for about 2/3 of the total annual rainfall in Niamey. The July–August mean number of events increased by about 25% between the period 1930–1949 and the period 1950–1969 (Fig. 14). It decreased by more than 1/3 in the years 1970–1989 compared with the 1950–1969 average. Thus, the decrease in the number of rainfall events is responsible for a rainfall deficit of about 2/9 between the period 1950–1969 and the period 1970–1989, supposing (as is justified by the curves of Fig. 13) that the mean event rainfall remained unchanged. Since the overall
rainfall deficit is close to 25% (see Section 4) we are led to conclude that almost 90% of the rainfall deficit of the 1970–1989 years in comparison with the 1950–1969 high, has been caused by the diminution of the number of rainfall events during the core of the rainy season. Similarly the explanation of the 1950–1969 high (at least in terms of rainfall distribution) is found in the increase of the mean number of events in July and August.

Combining the analysis of the LD1 and LD2 curves, the rainy season in Niamey may be divided into three stages, the characteristics of which are given in Table 4. All the figures of this table have been crudely rounded in order to give the order of magnitudes of the contribution of each stage to the seasonal rainfall. The proportion of the July–August rainfall to the seasonal total is slightly reduced during the dry years, which means that the rainfall deficit is larger during the core of the rainy season than on the margins. Whereas it appears possible to compute meaningful statistics on the basis of calendar months for Niamey, this may not hold for stations at higher latitudes, where the limits between the three parts of the rainy season could fall in the middle of months.

Table 4
Contribution of each phase of the rainy season to the total rainfall in Niamey. Rainfall values are in mm and the total is also given in percentage of the annual rainfall

<table>
<thead>
<tr>
<th></th>
<th></th>
<th></th>
<th></th>
<th></th>
</tr>
</thead>
<tbody>
<tr>
<td>May–June</td>
<td>11 ± 1.5</td>
<td>10 ± 1.5</td>
<td>10 ± 1.5</td>
<td>10 ± 1.5</td>
</tr>
<tr>
<td>Mean number of</td>
<td>11</td>
<td>9</td>
<td>11</td>
<td>9</td>
</tr>
<tr>
<td>Total rainfall</td>
<td>100 (18%)</td>
<td>100 (18%)</td>
<td>110 (17%)</td>
<td>95 (19%)</td>
</tr>
<tr>
<td>July–August</td>
<td>14 ± 2</td>
<td>11.5 ± 1.5</td>
<td>11.5 ± 1.5</td>
<td>11.5 ± 1.5</td>
</tr>
<tr>
<td>Mean number of</td>
<td>27 ± 3</td>
<td>30 ± 3</td>
<td>37 ± 4</td>
<td>26 ± 3</td>
</tr>
<tr>
<td>Total rainfall</td>
<td>380 (67%)</td>
<td>350 (63%)</td>
<td>425 (65%)</td>
<td>300 (61%)</td>
</tr>
<tr>
<td>September</td>
<td>9.5 ± 1.5</td>
<td>11 ± 1.5</td>
<td>10 ± 1.5</td>
<td>10 ± 1.5</td>
</tr>
<tr>
<td>Mean number of</td>
<td>8</td>
<td>9</td>
<td>10</td>
<td>8</td>
</tr>
<tr>
<td>Total rainfall</td>
<td>75 (13%)</td>
<td>100 (18%)</td>
<td>100 (18%)</td>
<td>80 (17%)</td>
</tr>
<tr>
<td>April and October</td>
<td>10 (2%)</td>
<td>10 (2%)</td>
<td>15 (3%)</td>
<td>15 (3%)</td>
</tr>
<tr>
<td>Total rainfall</td>
<td>565</td>
<td>560</td>
<td>650</td>
<td>490</td>
</tr>
</tbody>
</table>

7. Validation of the model

A major asset of the leak distribution as a model of rainfall regimes is that once it has been calibrated for a given time step \( T \), the c.d.f. of the rainfall accumulated over \( k \) time steps is still a leak distribution, assuming that an average scale parameter LD1 can be computed over the duration \( kT \). When daily data are available it is by far preferable to choose \( T = 1 \) day, rather than to 10 or 30 days, in order to minimise the correlation between the parameters of the leak distribution, as stated in Section 3.2. A straightforward way of independently validating the approach proposed in Section 5 is thus to compare, for a given duration of rainfall accumulation \( kT \), the theoretical distribution derived from the daily LD parameter curves, LD1(\( t \)) and LD2(\( t \)), and the experimental distribution. This will be done below for the August daily (\( k = 1 \)) and 10-day (\( k = 10 \)) rainfall distributions in Niamey, for two of the subperiods studied in Section 6, that is 1950–1969 and 1970–1989.

A monthly average of the LD1 parameter is first computed as a weighted mean of the 31 daily LD1 parameters of August:

\[
LD_{1,\text{Aug}} = \frac{1}{243} \sum_{i=213}^{243} LD_1(i)LD_2(i)
\]

(30)

where 213 and 243 are the first and last days of August, expressed as Day of year.

The model of the average daily (\( k = 1 \)) rainfall distribution for the month of August is defined by the two following parameters:

\[
LD_{1}^{(k=1)} = LD_{1,\text{Aug}}
\]

(31)

while the model of the average 10-day (\( k = 10 \)) rainfall distribution is defined by the two following parameters:

\[
LD_{1}^{(k=10)} = LD_{1,\text{Aug}}
\]

(32)

The theoretical distributions obtained from Eqs. (21)–(24) are compared with the experimental distributions of observed rainfall in Fig. 15. The correspondence is excellent for all four examples (daily and 10-day time steps for the wet period 1950–1969 and the dry period 1970–1989). Given that the LD curve (full line) was not fitted to the plotted data (dots), the similarity of the observed and computed distributions constitute a strong validation of the approach proposed here. The possibility of describing the rainfall distributions at various time steps by a simple two-parameter model is thus demonstrated for the Sahelian rainfall.

Note that a different set of parameters could be computed for each of the three 10-day
L. Le Barbé, T. Lebell


periods (that is d1 for 1–10, d2 for 11–20 and d3 for 21–31), modifying Eq. (33) as:

\[
\begin{align*}
\text{LD}_1^{(d1)} &= \text{LD}_1^{(d1)} \\
\text{LD}_2^{(d1)} &= \sum_{i=2}^{10} \text{LD}_2^{(d0)}
\end{align*}
\]

(34)

for the first 10-day period of August (i would range from 11 to 20 for d2 and from 21 to 31 for d3).

However, the fit of the distributions plotted in Fig. 15 to the observations is sufficiently good to assume a single 10-day distribution for all three 10-day periods of August in Niamey. The same does not necessarily hold for any other month or any other Sahelian station. In such cases the versatility of the leak distribution allows the identification of the proper set of parameters for each 10-day (or 5-day for that matter) period. It consequently provides an easy and coherent method to detect significant differences in the distribution of 10-day periods belonging to the same month.

8. Conclusion

The rainfall of the past 4 decades in the Sahel is distinguishable by the succession of 2 wet decades (1950–1969) and 2 dry decades (1970–1989), a phenomenon reminding us of the Joseph effect already noted by Mandelbrot and Wallis (1968). This paper has attempted to characterise the rainfall climatology for each of these two periods over Central Sahel. To that aim, a relatively simple stochastic model, belonging to the compound Poisson processes family was used. This model, known as the leak distribution, allows the decomposition of the rainfall fluctuations into two terms: the fluctuations of the mean event rainfall, and that of the mean number of rainfall events over any period of accumulation. The model proves extremely well suited to the description of the Sahelian rainfall at a regional scale, providing a coherent set of parameters whatever the time scale considered between 1 day and 1 month.

A remarkable spatial coherence emerges from this analysis, regarding the changes associated with the drought. All over the area of study (first Niger as a whole, then the 5° × 4° region centred on the HAPEX-Sahel square), the decrease of rainfall is closely linked to a decrease in the number of rainfall events, except in the extreme south. This decrease is especially important for the core of the rainy season (July, August), but in the north (14 to 15°N) it is observed over the whole rainy season. In contrast, the mean event rainfall varies little. Thus the main change that explains the rainfall deficit of the recent decades clearly appears to be the number of rainfall events rather than the average strength of the rainy events or the shortening of the rainy season.

Fig. 15. Daily (top), 10-day (middle) and monthly (bottom) rainfall distribution in Niamey for the periods 1950–1969 (left) and 1970–1989 (right). The model (full line) is a leak distribution (LD) whose parameters were computed by averaging the daily parameters of Figs. 13 and 14 for the corresponding periods. Dots are the observed rainfall. Although not fitted directly to these observed rainfalls the LD model closely reproduces the observed distribution.
The fact that the length of the rainy season did not change between the wet and dry periods supports the idea that the drought is not primarily linked to a shift in the average position of the ITCZ. There is a need to identify the factor responsible for the triggering of convection (which may be a consequence of the general atmospheric circulation or of local conditions), which will provide a physical basis for understanding the diminution in the number of Sahelian storms.

The consequences of this diminution in the number of rainy events, spread over a period whose length did not vary, are important for agriculture and hydrology. For instance, research on millet should probably not concentrate only on improving varieties with a short vegetative cycle but on the development of new varieties resistant to dry spells during the vegetative cycle as well. Also, the increase of the average duration between two storms in the core of the rainy season is likely to reduce the proportion of rainfall lost as surface runoff, the mean rainfall by storm being equal, because the soil has more time to dry out. This research on the rainfall regimes of West Africa is presently continuing so that it can be extended to the whole of the Sahel.

References