ABSTRACT

Because of a lack of data, the structural characteristics of rain fields over the sea are poorly documented. Coastal radars offer an opportunity to observe the distribution of rain parameters at the land–sea transition. In this study, two datasets on rain fields collected over the Atlantic coast: one at midlatitude, in a westerly general atmospheric circulation, that is, onshore, in the southwest of France, the other at tropical latitude, in an easterly general atmospheric circulation, that is, offshore, in the west of Senegal (Africa), are analyzed. In the two areas, the rain volume, or cumulative rainfall, is found to be markedly larger over land than over sea. However this difference is due mainly to a higher rain occurrence and duration over land than over sea. The mean rain rate, when raining, is almost the same over land and over sea at midlatitude and at tropical latitude. In addition, the mean rain rate is found to be constant through rain fields in which strong gradients of cumulative rainfall are observed [when no other kind of forcing (not considered in this work), notably orographic, is present]. Thus, for example, in Senegal a meridional change of the annual cumulative rainfall from 300 to 1200 mm over 400 km is not associated with any mean rain-rate variation. In a similar way, the parameters of the probability density function of the rain rate and the distribution of the statistical variation coefficient are not influenced by the gradient of the cumulative rainfall. The studied rain fields are thus approximately ergodic. These results strengthen the validity of the probability density function of rain rate as a climatic characteristic of rain fields and the reliability of the statistical method of estimating the average rainfall by area integrals over large space scales and timescales. The variation coefficient of the rain rate is found to be constant and close to 2.24 over sea and over land for mid- and tropical latitudes.

1. Introduction

Figure 1 presents the distribution of the cumulative rainfall in the southwest of France, east of the Atlantic coast of the Bay of Biscay, and in Senegal, on the western African coast. As in most similar rain field representations, isohyets end at the coast. They are not given over the sea because the requisite data (usually the ground network rain gauge data) are not available. Thus the rain field characteristics over the sea are poorly documented.

Several studies about the rain field over the sea, far from the coast, were performed. Examples include the Global Atmospheric Research Program’s Atlantic Tropical Experiment (e.g., Hudlow 1979) and the Tropical Ocean and Global Atmosphere Coupled Ocean–Atmosphere Response Experiment (e.g., Webster and Lukas 1992). The results show the influence of the sea surface temperature on the spatiotemporal rainfall distribution.

They also display the differences between the structure and organization of rain fields over sea and over land. The rain field distributions are notably more even and smoother over sea than over land. The microphysical and dynamical processes responsible for the precipitation development are clearly qualitatively the same over sea and over land. However, some by-products of the deep convective clouds are very different over sea and over land, such as the occurrence of lightning, which is much higher over land than over sea (Zipser 1994; Christian et al. 1999; Boccippio et al. 2000; Seity et al. 2001; see also http://thunder.nsstc.nasa.gov/data/). Moreover, hail size is observed to be smaller in coastal areas than inland (Dessens et al. 2000; Vinet 2001). These differences are attributed to a lesser maximal intensity of convection over sea than over land, because the uniformity of the surface over the sea leads to the release of the convective instability in the form of a more homogeneous field, that is, with lesser maximal values (Williams et al. 1992; Zipser 1994; Zipser and Lutz 1994; Mesnard and Sauvageot 2001, manuscript submitted to J. Geophys. Res.).

The object of this paper is to describe and to discuss the characteristics of the rainfall distribution in coastal areas, which offers an opportunity to observe the land–sea contrast. Is it possible to observe sea–land differ-
for the kind of land–sea comparison addressed in this paper. In comparison, the space sensors, infrared and microwave alike, have a lesser space–time resolution and accuracy for rain-rate retrieval (e.g., McCollum et al. 2000). Recent results on tropical rainfall distributions from the Tropical Rainfall Measuring Mission (TRMM) combined with other satellite data are very promising; however, the time resolution is still crude (Adler et al. 2000).

This paper is concerned with the distribution of the cumulative rainfall, rain duration, and mean and variance of the rain rate. Two observational sites are considered, one at midlatitude with a landward general atmospheric circulation and the other at tropical latitude with a seaward circulation.

### 2. Data

The data were collected with the radars of the meteorological stations of Bordeaux–Mérignac (44°52′N, 0°30′W, altitude 70 m) in southwestern France and Dakar–Yoff (14°34′N, 17°29′W, altitude 30 m) in Senegal (Fig. 1). The technical characteristics of these radars are given in Table 1.

The climate of the Bordeaux area is of oceanic type. Most of the rainy events are zonal frontal cyclonic disturbances, that is, moving eastward, from sea toward land. The climatic conditions are mostly homogeneous, although, during the summer season, the frontal systems are often reduced to the convective line associated with the cold-frontal line. The mean annual rainfall is about 800 mm distributed throughout the year. The Bordeaux area is flat; within a radius of 150 km around the radar, the altitude is below 200 m, except to the northeast, between azimuths 40° and 80°, at a distance of more than 130 km, where there are the first hills of the Massif Central, with an altitude below 400 m. About 200 km south of Bordeaux is the Pyrénées range, which is the border between France and Spain. Westward, the Pyrénées range is continued by the Cordillera Cantabriques, a line of mountains with an altitude below 3000 m, forming the southern limit of the Bay of Biscay. As clearly illustrated by Figure 1a, this orography is associated with a meridional gradient of the cumulative rainfall over the southwest of France, with values increasing up to 1600 mm over the Pyrénées range. Moreover, radar observations show that deep convective clouds starting over the Cordillera Cantabriques, an area of strong convective activity, frequently are carried...
away by the southwesterly flow, over the Bay of Biscay, up to the French coast. The radar is located 36 km east of the coast, which is almost straight and is directed north–south.

The climate of the Dakar area is of sahélien type. The rainy season is reduced to about 3 months, from early July to late September when the latitude of the intertropical convergence zone (ITCZ) is higher than 13°N. Most rainfalls become weaker and then disappear when crossing the coast and moving over the nearby ocean; inversely, a few systems grow stronger, advance over the sea, and seem eventually able to play a role in the genesis of the hurricanes of the west tropical Atlantic (Gray and Landsea 1992). The land east of Dakar is flat over a distance of longer than 500 km (altitude below 200 m). The mean annual cumulative rainfall (Fig. 1b) displays a strong meridional gradient, from 300, at the latitude of Saint Louis (16°01’N) near the Mauritanian border, to 1500 mm at Cap Skirring (12°20’N) near the Guinea–Bissau border, which is 400 km away. This gradient is associated with the meridional dynamic of the ITCZ. The radar is located at the airport of Dakar–Yoff. The coastline forms an angle of about 120° whose vertex, located in the west, is Dakar.

The characteristics of the dataset used in the study are given in Table 2. The radar of Bordeaux is an element of the French operational radar network managed by Météo-France. The data are collected continuously, with a scanning repetition period of 5 min. Only 4 months of the summer season were used so as to minimize the problem linked to the altitude of the precipitation melting layer (see below). The radar of Dakar is an operational radar dedicated to the observations of the Office National de la Météorologie du Sénégal and the Association pour la Sécurité de la Navigation Aérienne (ASECNA); however, it is activated only during rainy events. The data acquisition was performed manually by the staff of the Laboratoire de Physique de l’Atmosphère of the University Cheikh Anta Diop of Dakar, using a “Sanaga” acquiring system (Sauvageot and Despau 1990). The scanning repetition period is between 10 and 20 min. Seven rainy seasons (1993–99) were used. The reason for the difference between the time size of the Bordeaux and Dakar datasets is that in the Dakar area about 72% of the total annual precipitation falls in less than 8 h (because of the high mean rain rate of the convective part of the sahélien squall lines; see section 3). For the north of the studied area (Saint Louis) this duration is reduced to less than 4 h which is very short for a statistical study. Using 7 years of data enables us to obtain, for Dakar, a dataset with a time size similar to the one of Bordeaux.

The minimal rain rate detectable with a radar (MDRR) increases with distance. Thus, to avoid a bias in the data homogeneity, only reflectivities higher than the value of MDDR at 200 km were taken into account. This value is 15 dBZ at Bordeaux, which corresponds approximately to rain rate $R > 0.2$ mm h⁻¹, and is 18 dBZ, or $R > 0.3$ mm h⁻¹, at Dakar.

The two radars are regularly calibrated (in the electronic sense) by their maintenance staff to provide correct values of the radar reflectivity factor $Z$. The $Z$ values were converted to $R$ using a $Z$–$R$ relation of the usual form: $Z = aR^b$ where $a$ and $b$ are coefficients depending mainly on the drop size distribution (DSD), if $Z$ is assumed to be correct (e.g., Joss and Waldvogel 1990). These coefficients were determined using the probability matching method (PMM) proposed by Calheiros and Zawadzki (1987). PMM leads to the computation of corrected values (in the statistical sense) of $a$ and $b$ as a function of $r$, the radial distance from the radar, that is, taking into account the differences of reflectivity between the precipitation observed aloft by the radar (whose beam rises above the ground with distance) and the rain at ground level.

The parameters of the probability density function of

### Table 2: Characteristics of the dataset. CAPPi is constant-altitude plan position indicator, PPI is plan position indicator, $a$ is angle elevation, $r$ is radar–target distance, and $Z$ is radar reflectivity factor.

<table>
<thead>
<tr>
<th>Location of radar</th>
<th>Bordeaux–Mérignac</th>
<th>Dakar–Yoff</th>
</tr>
</thead>
<tbody>
<tr>
<td>Scanning mode</td>
<td>CAPPI with $a = 1.5^a$ for $r &lt; 50$ km, $a = 0.4^b$ for $r &gt; 50$ km</td>
<td>PPI, $a = 0.8^c$</td>
</tr>
<tr>
<td>No. of scans</td>
<td>10 052</td>
<td>7407</td>
</tr>
<tr>
<td>Sampling interval</td>
<td>5 min</td>
<td>Between 10 and 20 min</td>
</tr>
<tr>
<td>Pixel size</td>
<td>$1 \times 1$ km²</td>
<td>$1 \times 1$ km²</td>
</tr>
<tr>
<td>No. of steps for $Z$ coding</td>
<td>52</td>
<td>256</td>
</tr>
</tbody>
</table>

### Table 3: Coefficients of the $Z$–$R$ relations computed from the PMM method and used to obtain the field of rain rate. Here, $r$ is the radar–target distance (km).

<table>
<thead>
<tr>
<th>Regions</th>
<th>Bordeaux</th>
<th>Dakar</th>
</tr>
</thead>
<tbody>
<tr>
<td>Linear coefficient</td>
<td>$a = 278e^{-0.0145r}$</td>
<td>$a = 383e^{-0.0041r}$</td>
</tr>
<tr>
<td>$40 \leq r \leq 90$ km</td>
<td>$60 \leq r \leq 180$ km</td>
<td></td>
</tr>
<tr>
<td>Exponential coefficient</td>
<td>$b = 1.46 + 0.0013r$</td>
<td>$b = 1.24 - 0.0017r$</td>
</tr>
</tbody>
</table>
R, \( P(R) \), needed to implement PMM (see Atlas et al. 1990) were deduced from several years of disdrometer observations of DSDs in southwestern France (Sauva-geot 1994) and in Dakar (1997–99). The parameters of the probability density function of \( Z, P(Z) \), to be associated with \( P(R) \) in PMM were deduced only from the radar data observed over land. The values of \( a(r) \) and \( b(r) \) computed in this study are given in Table 3.

To check the correctness of the PMM calibration procedure, the radar-determined cumulative rainfalls \( \Sigma \ H \) were compared with the measurements of the ground rain gauge network \( \Sigma \ G \) (Wilson 1970). The observed differences \( E = \Sigma \ G / \Sigma \ H \), inside the area of radius 100 km from the radar, were smaller than 20% for the individual rainy events.

This calibration performed from the data observed over land was also used for the oversea area, because there is no reason to suspect important differences between DSDs observed over land and over sea for the same climatic area. DSD microstructure depends mainly on falling-drop interactions and not on the initial DSD (e.g. Srivastava 1978). Yet more research and data collection would be useful to reach definitive conclusions about the comparison of land–sea DSDs. However, the available literature concerning the rain microstructure does not suggest any systematic sea–land difference in DSD (Hudlow 1979; Joss and Waldvogel 1990; among others). For example Sauva-geot and Lacaux (1995) found almost the same \( Z-R \) coefficients for continental Africa and for a coastal site of equatorial Africa. It can also be considered that if significant land–sea DSD differences exist in the two sites of this study, there would appear significant land–sea differences in the \( Z \) or \( R \) land–sea distribution. Section 3 will show that such differences are not observed. On the other hand, as will be seen in section 3c, this calibration method does not correct for the effect of the truncation of the \( R \) distribution (that is the MDRR) on the average and variance of the rain rate.

The parameters of the rain field discussed in the following section are presented up to a distance of 200 km. At such a distance, the upper edge of the radar beam for plan position indicator observation, with minimal elevation angle (see Table 2), is at an altitude of 4.5 km (2.25 km at 100 km) at Bordeaux and 5 km (2.5 km at 100 km) at Dakar. In the Bordeaux area, the mean monthly altitude of the melting layer is always lower than 4.5 km of altitude; it is below 2.25 km during the five coldest months of the year (Bourrel et al. 1994). The rain-rate distribution also includes a significant component of stratiform rain generated by clouds of weak thickness, with top in the lower troposphere (see section 3). That is why, in the Bordeaux area, only the four summer months are used. In spite of that, at a distance beyond 100–120 km from the radar, because of the altitude of the radar beam, part of the stratiform rain, even with \( R \) higher than the MDRR, is not detected. Thus the radar data samples suffer an underweighting, or filtering, for stratiform rain, and thus the thick convective rain is overweighted. In section 3, this filtering is shown to affect the distribution of some rain parameters beyond about 120 km. At Dakar, most of the rain systems are associated with deep convective clouds such as squall lines, in which even the stratiform rain clouds have their top near the tropopause (Johnson et al. 1990). Besides, during the rainy season, the melting layer is above 4500 m of altitude. The result is that the stratiform precipitation filtering effect is not observed for distance shorter than about 180 km. However, the rain-parameter distributions are presented up to 200 km for both sites in such a way that the area of validity is apparent.

Moreover, screening effects are observed at Bordeaux for azimuths 300° and 310° and at Dakar for azimuths 190° and 210°. The cause is, for Bordeaux, a building and, at Dakar, two small old volcanic hills, named the Mamelles, close to the radar.

Close around the radar, the meteorological signal usually is blurred by the ground (or surface) clutter echoes in such a way that corresponding data are not usable. Of course, in this study, the surface clutter–biased areas are not considered for the estimate of average values but are maintained in all the figures showing the rain parameter distributions because their arrangement is in-

<table>
<thead>
<tr>
<th>Sea</th>
<th>Land</th>
<th>North</th>
<th>South</th>
<th>Sea</th>
<th>Land</th>
<th>North</th>
<th>South</th>
</tr>
</thead>
<tbody>
<tr>
<td>( \langle H \rangle ) (mm)</td>
<td>120</td>
<td>159</td>
<td>122</td>
<td>161</td>
<td>298</td>
<td>632</td>
<td>344</td>
</tr>
<tr>
<td>( \langle T \rangle ) (h)</td>
<td>106</td>
<td>133</td>
<td>111</td>
<td>128</td>
<td>53.4</td>
<td>105.0</td>
<td>56.7</td>
</tr>
<tr>
<td>( \langle \sigma_R \rangle ) (mm h(^{-1}))</td>
<td>1.13</td>
<td>1.20</td>
<td>1.10</td>
<td>1.26</td>
<td>5.58</td>
<td>6.02</td>
<td>6.06</td>
</tr>
<tr>
<td>( \langle \mu_R \rangle ) (mm h(^{-1}))</td>
<td>1.50</td>
<td>1.89</td>
<td>1.47</td>
<td>1.95</td>
<td>10.36</td>
<td>10.85</td>
<td>10.61</td>
</tr>
<tr>
<td>( \langle \sigma_U \rangle ) (mm h(^{-1}))</td>
<td>0.66</td>
<td>0.67</td>
<td>0.59</td>
<td>0.72</td>
<td>4.80</td>
<td>5.35</td>
<td>5.18</td>
</tr>
<tr>
<td>CV</td>
<td>1.37</td>
<td>1.45</td>
<td>1.31</td>
<td>1.63</td>
<td>11.61</td>
<td>11.88</td>
<td>11.41</td>
</tr>
<tr>
<td>CV</td>
<td>2.05</td>
<td>2.17</td>
<td>2.24</td>
<td>2.27</td>
<td>2.42</td>
<td>2.22</td>
<td>2.20</td>
</tr>
</tbody>
</table>
teresting to observe. The pixels concerned are clearly identifiable.

The areas used for the average parameter estimates given in Table 4, whose usefulness is discussed in section 3, were drawn to take into account the topographic particularities of the observed areas and to avoid a bias in the land–sea or north–south comparison by a fault in symmetry (Fig. 2).

Azimuthal scans with a low elevation angle are sometimes biased by anomalous radar propagations, usually termed “anaprop,” that modify the rain field observation and add surface echoes. In this work, the scans including anaprops were removed after a careful examination by eye of the whole dataset.

In this paper, time average of rain rate inside the individual radar pixels is written $\bar{R}$, and the space average of the individual pixel values over an area is written $\langle \bar{R} \rangle$.

3. Results

a. Cumulative rainfall

Figure 3 displays the field of radar-determined cumulative rainfall height $H$ (mm) for the two sites. For Bordeaux (Fig. 3a), beyond a radial distance of about 120 km, $H$ decreases because of the instrumental filtering effect stressed in section 2. The nonfiltering observation area is a circle of radius 100–120 km around the radar.

The north–south gradient of $H$ is found again on Fig. 3a, but, in addition, a clear sea–land gradient appears. The combination of these two gradients results in the slanted southwest–northeast orientation of the isohyets where they cross the coastline. As a whole, the cumulative rainfalls are higher over land than over sea, because $H$ values are 120 and 159 mm over sea and over land, respectively, which is a 33% difference (Table 4). The north–south values are 122 and 161 mm, respectively, which is a 32% difference.

For Dakar, the distribution of $H$ (Fig. 3b) leads to similar conclusions as at Bordeaux. The combination of a very strong north–south gradient ($H$ is about 5 times higher south of the studied area than north of it) and a sea–land gradient of $H$, resulting in the southwest–northeast direction of some isohyets (notably the 400-mm one), is also observed. The averaged values are 298 and 632 mm for sea and land, respectively, which is a difference of 112%. The north–south difference is 69% (Table 4). In Fig. 3b, any decrease of $H$ as a function of the radial distance is observed within the limit of the observed area, that is 200 km.

At Bordeaux, the surface echoes are shrouded in the rain-echo distribution. At Dakar, the surface-echo area is very asymmetrical and is mainly over the cape of Dakar, up to a distance of 60 km. Over the sea, the surface-echo area is circle-shaped with a radius of about 15 km, because the sea surface behaves as a specular reflector with low backscattering.

b. Rainfall duration

The distribution of the rainfall duration $T$ is presented in Fig. 4. For Bordeaux, $T$ is the total duration of rainfall above the MDRR during the sampling period of 4 months. For Dakar, $T$ is the corresponding mean annual value. The distribution of $T$ is similar to that of $H$. At
Fig. 3. Distribution of the cumulative rainfall: (a) Bordeaux area for the 4 summer months of 1996 and (b) Dakar area in annual mean.
Fig. 4. Same as Fig. 3, but for the rain duration (days).
FIG. 5. Probability density function of the conditional rain rate observed in the Bordeaux and Dakar areas over sea, land, northern, and southern parts of the fields.

Bordeaux (Fig. 4a) a strong decrease of $T$ is observed for $r$ greater than 120 km because of the bias sampling, but the areas of high $H$ values appearing in Fig. 3b beyond 120 km southwest and northeast of the radar are smoothed in Fig. 4a, because, as pointed out in section 2, the bias sampling for long distances favors the observation of short-duration deep-convective rain. The sea–land difference in duration (Table 4) is 25%, which is close to, although slightly lower than, the $H$ difference. The north–south difference is 15%, which is lower than the $H$ difference, showing that the rain rate is slightly higher in the south part (cf. section 3d).

For the Dakar area, the distribution of $T$ shows minimal values in the northwest quadrant and maximal values in the southeast one. The sea–land difference in rain duration is 97%, which is slightly smaller than the $H$ difference, suggesting that the rain rate is a bit higher on land. The north–south difference for $T$ is 72%, which is almost the same as the $H$ difference.

Surface clutter areas are prominently displayed on the $T$ distribution because they are permanent. At Dakar, over sea, the surface clutter area is at the same level of duration as over land.

c. Probability density function of the rain rate

The probability density function of the conditional rain rate (that is, when raining) or $P(R)$, for the various areas in discussion, is drawn in Fig. 5. These curves show that, at Bordeaux, the intermediate $R$ values have a slightly higher frequency over sea than over land, which is compatible with a more homogeneous convection over sea. At Dakar, for all the $R$ values higher than the mode of $P(R)$, the frequency is lower over sea than over land; it is, of course, the reverse for $R$ smaller than the mode. All that suggests a convection slightly less vigorous over sea than over land, as expected.

The north and south $P(R)$ curves for the Dakar area coincide almost exactly despite the strong gradient of the cumulative rainfall. A slight difference appears be-
between the north and south curves of Bordeaux, showing that the $R$ values are a bit higher in the southern part; a tentative explanation of that could be searched for in the cause of the north–south gradient of $R$, which, in the southwest of France, is associated with an orographic effect, which implies that the dynamical and microphysical processes of rain development change, in part, from north to south.

Figure 5 shows that the shape of $P(R)$ is compatible with a lognormal distribution, as generally accepted (Atlas et al. 1990; Kedem et al. 1990; Sauvageot 1994). A lognormal distribution is defined by two parameters, namely, the average and the variance of the variable. Figure 5 also shows the truncation of the left part of the observed distribution by the MDRR limitation. This truncation is much more severe for the Bordeaux than for the Dakar distribution. To quantify the effect of the truncation, the $P(R)$ parameters were corrected for truncation using the method described by Aitchison and Brown (1966) and Cohen (1959, 1991). Simulation of truncation from theoretically known $P(R)$ of a similar shape as in Fig. 5 was performed to test the efficiency of the used correction procedure. The tests show that this correction enables the retrieval of the untruncated lognormal parameters.

Table 4 gives the area-averaged value of the mean and standard deviation of $R$ computed directly from the observed raw data, that is, $\langle R \rangle$ and $\langle \sigma_{UR} \rangle$ (where subscript $U$ stands for uncorrected), respectively, and the truncation-corrected values, referred to as $\langle \mu_R \rangle$ and $\langle \sigma_R \rangle$. These last two parameters were obtained as follows. From the temporal series of the calibrated $R$ values, the parameters of the probability distribution function (pdf) of $R$, assumed to be lognormal, were computed for each pixel and were corrected for truncation. Then, the corrected values of $\mu_R$ and $\sigma_R$ were averaged over the various areas.

What appears is that the truncation strongly modifies the mean value of $R$ for the Bordeaux area, because the corrected value is about one-half of the observed $\langle R \rangle$, for the four areas of Fig. 2. For the Dakar area, the corrected values are only 14% lower than the observed ones, in average for the four areas.

The parameter $\langle \mu_R \rangle$ for the four areas is 0.66 and 5.09 for Bordeaux and Dakar respectively. These values are perfectly compatible with the values of $\mu_R$ found for midlatitude (between 0.63 and 0.95 mm h$^{-1}$) and for Sahelian Africa (5.14 mm h$^{-1}$) by Sauvageot (1994). The scattering of $\langle \sigma_R \rangle$ in Table 4 is also small.

d. Average rain rate

The distribution of the corrected average rain rate $\mu_R$ is shown in Fig. 6 for the two areas.

In the Bordeaux area (Fig. 6a), for $r$ less than 110 km from the radar, the $\mu_R$ distribution is mostly homogeneous. The sea–land difference for the averaged $\mu_R$ value is less than 2%, and the north–south difference is 22%, which is almost constant around 0.66 mm h$^{-1}$ (Table 4).

Beyond about 120 km, the $\mu_R$ distribution is more confused and increases because of the low-cloud filtering effect. Some cores of high value, partly associated with the orography, appear. Screening effects are also observed in azimuths 300° and 310°.

In the Dakar area, the $\mu_R$ values are much higher than in the southwest of France, by a factor of about 7.7. However, the $\mu_R$ distribution (Fig. 6b) is also mostly homogeneous for $r$ less than 180 km and except for the northwestern quarter plan beyond 100 km (see below). The area-average $\mu_R$ (Fig. 2 and Table 4) are 4.80 and 5.35 mm h$^{-1}$ for sea and land, respectively, which is a difference of 11%. The north–south values are 5.18 and 5.03 mm h$^{-1}$, respectively, which is a difference of less than 3%. The global average is 5.09 mm h$^{-1}$. To sum up the characteristics of the $\mu_R$ distribution, what appears is that $\mu_R$ is almost constant for the whole observed area, with slightly lower values over the sea.

In the northwestern quarter plan, beyond about 100 km offshore, high values of $\mu_R$ (higher than 10 mm h$^{-1}$) are observed in an area for which rain durations (Fig. 4b) are low. Careful examination of the data shows that these high $\mu_R$ values are associated with a very low number of events in which, contrary to what is observed in most of the cases, the squall line does not disappear over sea but grows stronger, maybe on account of a tropical cyclogenesis process.

At Bordeaux, the surface echoes (ground clutter) are partly shrouded inside the rain distribution because the level of their averaged intensity is similar to the local mean rain rate. It is the same at Dakar, except that, over the sea, the area affected by the surface echoes displays a lower $R$ value. In fact, when it does not rain over this area, pixels with a radar reflectivity factor larger than the MDRR but lower than the local mean $R$ value continue to be detected and thus are taken into account in the $R$ average.

e. Standard deviation

The distribution of the corrected standard deviation of $R$, $\sigma_R$, is shown in Fig. 7, and the averaged values of $\langle \sigma_R \rangle$ are given in Table 4. It is seen that $\sigma_R$ is much higher at Dakar than at Bordeaux. For both the Bordeaux and Dakar areas, $\sigma_R$ may be slightly lower over sea than over land, by 6% and 3%, respectively. There is the same difference between the north and south areas at Bordeaux, and no difference is seen at Dakar. It is suggested that the lower sea $\sigma_R$ value at Bordeaux mirrors the differences in the convection distribution, which is assumed to be more homogeneous over sea than over land, as indicated in section 1.

f. Variation coefficient

In order to assess the relative space variation of $\sigma_R$ and $\mu_R$, the field of the variation coefficient CV, $\sigma_R/\mu_R$. 
Fig. 6. Same as Fig. 3, but for the conditional average rain rate corrected for truncation, μₙ.
Fig. 7. Same as Fig. 3, but for the std dev of the conditional rain rate, \( \sigma_x \).
of the distribution is presented in Fig. 8, and the average values of CV are given in Table 4. For comparison, the uncorrected variation coefficient, that is, $CV_U = (R)/\sigma_{t\bar{R}}$, is also given in Table 4.

Sauvageot (1994) found that $\sigma_R$ and $\mu_R$ values, when averaged over large space and time samples, are linked by the relation $\sigma_R^2 = 5\mu_R$, that is, CV constant and equal to about 2.24, and that this relation stands up over a wide range of rain-rate variation, including the mid- and tropical-latitude rain rates. However, this property was not discussed for the small-scale variations of CV. Figure 8 and Table 4 show that, for mid- and tropical-latitude areas, the values of CV, both for the sea–land and north–south regions, are weakly scattered and very close to the value proposed by Sauvageot (1994), given that the mean values are 2.18 and 2.27 for Bordeaux and Dakar, respectively. The mean value for Bordeaux and Dakar is 2.23. Thus, it seems that the CV = 2.24 relation applies to the small scale. This result is interesting, because it shows that only the knowledge of the mean rain rate enables the definition of $\sigma_R$ and $P(R)$. It can also be noted that, in the field of CV (Fig. 8), the surface echoes and the effects of the screening on the azimuthal directions 300° and 310° at Bordeaux and 190° and 210° at Dakar appear strongly reduced. The reason is that the screening effects decrease both the standard deviation and the average in such a way that their ratio is less affected than the direct parameters. For similar reasons, the ground clutter is not prominent on the CV distribution.

The homogeneity and stability of the $\mu_R$ and $\sigma_R$ distributions imply that of $P(R)$. Such a homogeneity, through rain fields in which strong cumulative rainfall variations are observed, is interesting, because it shows that the observed rain fields are approximately ergodic [i.e., $\mu_R$ and $\sigma_R$ do not differ when computed over different data samples (e.g., Bendat and Piersol 1971)], and thus it justifies the use of $P(R)$ as a basic climatic characteristic of rain fields.

The stability of $P(R)$ implies, in turn, that of $S(\tau)$, the linear coefficient that links the average rain rate ($R$) to the fractional area where it is raining $F(\tau)$, that is, $(R) = S(\tau)F(\tau)$, as proposed by the area-integral (or threshold) method (Doneaud 1984). Indeed, $S(\tau)$ can be written as $S(\tau) = \int_0^\tau RP(R) dR/R$, $P(R) dR$, which shows that $S(\tau)$ is determined by $P(R)$ (e.g., Atlas et al. 1990; Sauvageot 1994; Sauvageot et al. 1999). The results show, thus, that $S(\tau)$ is not influenced by a gradient of cumulative rainfall. All that evidence strengthens the reliability of the area-integral method for area-average rain-rate estimation over large space scales and timescales, notably from space observations (e.g., Meneghini and Jones 1993; Meneghini 1998; Sauvageot et al. 1999).

4. Summary and conclusions

Two datasets were collected with radars located near the Atlantic coast, one at midlatitude (Bordeaux) with an eastward general circulation and the other at tropical latitude (Dakar) with a westward general circulation. The datasets were used to analyze comparatively the distribution of the rain parameters over sea and over land.

The main conclusions are first that the rain volume, or cumulative rainfall, is higher over land than over sea, by 33% and 112% for the midlatitude and tropical areas, respectively. However, the bulk of these differences is not due to variations of rain rate but of rain duration, which is longer over land than over sea, by 25% and 97% for mid- and tropical latitude, respectively.

The average rain rate is much stronger for tropical latitude (5.1 mm h⁻¹ for the corrected value) than for midlatitude (0.66 mm h⁻¹), but inside the same area it is homogeneous. At Bordeaux no sea–land difference is found. For Dakar, the sea–land difference is an increase of 11% over land. The results are compatible with a convection slightly less vigorous over sea than over land. The maximum $R$ values observed over sea are a bit lower than over land.

In the two sites, strong north–south gradients of cumulative rainfall are observed, by about 30% over 200 km in Bordeaux and by 500% over 400 km in Dakar. No average rain-rate variation is found to be associated with this gradient in Dakar, and only 22% (probably orographically induced) in Bordeaux. The standard deviation of the rain rate is also found to be homogeneous and not influenced by the cumulative rain-rate distribution.

The probability density distribution of the rain rate is well represented by a lognormal function, which is determined by two parameters, the mean $\mu_R$ and the standard deviation $\sigma_R$. Thus, the stability of $\mu_R$ and $\sigma_R$ through rain fields displaying a strong heterogeneity of the cumulative rainfall distribution, such as observed in this study, implies the same stability for the pdf of $R$ or $P(R)$. It shows that the studied rain fields are approximately ergodic and justifies the validity of $P(R)$ as a significant rain field characteristic. In a similar way, the stability of $P(R)$ implies the stability of $S(\tau)$, the linear coefficient of the relation between the average rain rate and the fractional area. The stability of $S(\tau)$ is fundamental to the rainfall estimate by the area-integral method, notably over large space scales and timescales from satellite observations.

At the scale of the radar-observed area, the variation coefficient of the rain rate, that is, $\sigma_R/\mu_R$, is also found to be homogeneous and very close to 2.24, the value proposed by Sauvageot (1994).

The results on the distribution of the mean and standard deviation of the rain rate suggest that the conditions underlying the forcing, development, and cycle of convective storms present some differences over land and over sea but are mostly homogeneous inside each domain and are notably weakly sensitive to meridional meteorological variations.

The distribution of rainfall parameters presented here lends itself to a comparison with the distribution of total flashes (cloud to ground and intracloud) as observed
Fig. 8. Same as Fig. 3, but for the variation coefficient.
with the National Aeronautics and Space Administration Optical Transient Detector and Lightning Imaging Sensor from space (e.g., Boccippio et al. 2000; http://thunder.nsstc.nasa.gov/data/). Such comparison is especially interesting for the Dakar area, where strong zonal and meridional gradients of the lightning activity are observed [the case of Bordeaux has been analyzed by Seity et al. (2001)]. The comparison for Dakar shows clearly that, over land, the occurrence of flashes is almost linearly related to the cumulative rainfall; the variation of the annual flash rate is about a factor of 5 between Saint Louis and the Guinea border, similar to \( H \). The land–sea variation of flash rate is much stronger, a kind of discontinuity, showing that the flash production of sea storms is very low, in agreement with Zipser (1994), Zipser and Lutz (1994), and Boccippio et al. (2000).

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