Influence of the Vertical Profile of Reflectivity on Radar-Estimated Rain Rates at Short Time Steps

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ABSTRACT

The present study aims to demonstrate the major influence of the vertical heterogeneity of rainfall on radar–rain gauge assessment. For this purpose, an experimental setup was deployed during the HYDROMET Integrated Radar Experiment (HIRE-98) based on a conventional S-band weather radar operating at long range (90 km), an X-band vertically pointing radar, and a network of 25 tipping-bucket rain gauges. After calibration and attenuation corrections, the X-band radar data enables the estimation of the vertical profile of reflectivity (VPR) time series. Screening and VPR correction factors are derived for the distant S-band radar measurements. The raw and corrected S-band radar estimates are compared to rain gauge measurements for various integration time steps (6–30 min). Considering about 12 h of intense Mediterranean precipitation, the VPR influence at the X-band radar site is clear for all the time steps considered. For instance, a continuous increase in the Nash efficiency for the corrected radar data compared to the rain gauge data (0.85 for the 6-min time step, up to 0.93 for the 30-min time step) is observed while this criterion remains less than 0.15 for the raw radar data, regardless of the time step. The effect of the low-level reflectivity enhancement on the radar–rain gauge assessment was also found to be very important in the considered configuration. The establishment of reliable VPR climatologies is therefore a challenge in order to better account for such effects that are not observable at long range from the radars. The spatial validity of the VPR correction derived from a point sensor like the vertically pointing radar is also investigated. As a result of the high space–time variability of rainfall, such a punctual VPR correction has an efficiency limited to areas of about 20 km² (200 km²) for the 6-min (30 min) integration time step.

1. Introduction

The vertical heterogeneity of rainfall can be considered as one of the dominant sources of error in radar measurements of rainfall (Joss and Waldvogel 1990). Since both the altitude and the size of the radar beam increase with range, this effect is likely to limit the effective range of radar for hydrological applications. In mountainous regions, the combination of vertical profile of reflectivity (VPR) and visibility problems related to the presence of the relief lead to even more serious limitations (Joss and Lee 1995; Pellarin et al. 2002). Since the early work of Smith (1986), several approaches have been proposed to cope with the VPR. These methods range from climatological corrections depending on seasons (Koistinen 1991; Joss and Lee 1995) or on geographical/meteorological considerations (Collier 1986) to attempts at estimating and correcting for the actual temporal and spatial variations of the VPR determined from volume-scan radar data. The latter path was followed for instance by Andrieu and Creutin (1995) who proposed a VPR identification technique based on an inverse approach. This method was first implemented with a two-elevation scanning radar. This relatively poor sampling allowed corrections at the hourly time step using a single VPR for all the considered domain (about 10 000 km²). These corrections proved to be effective but not to the extent that should have been expected (Creutin et al. 1997). Vignal et al. (1999) proposed a generalization of the VPR retrieval technique to full volume-scan radar data. Applying this method in the context of the Swiss and the U.S. radar networks allowed the determination of local estimates of the VPR for areas ranging from 50 up to 2000 km² (Vignal et
Fig. 1. Topography of the Marseille area, positions of the S-band (Nîmes) and X-band (Vernet and Vallon Dol) radar systems and of the rain gauges.

al. 2000; Vignal and Krajewski 2001). Vignal et al. (2000) compared the inverse method with the operational method of the Swiss radar network which takes into account the current weather conditions by estimating a mean vertical profile from volumetric radar data collected close to the radar site. It was found that both methods improve the radar–rain gauge assessment criteria very significantly in comparison with the raw and the climatological VPR-corrected radar data at the hourly time step. However, the local VPR approach provided only a marginal improvement over the mean profile approach, a rather disappointing result owing to the difference in the computational resources required for the two approaches. This result may be due to instrumental limitations related to the scanning strategy of the radar: for instance, the lower part of the VPR becomes less and less determinable as the range from the radar increases. It may also be due to the nonvalidity of the required assumption on the spatial homogeneity of the VPR over the considered subdomains during the considered time step (1 h in the mentioned references).

The present paper aims at documenting such “high-frequency” variations of the VPR, that is, variations at time steps less than 1 h and at quasi-punctual scales, and their impact in terms of radar rainfall estimation for a radar operating at long range from the region of interest. The work is based on a dataset collected during the HYDROMET Integrated Radar Experiment (HIRE-98), which took place in Marseille in 1998 (Uijlenhoet et al. 1999). Marseille, like the entire French Mediterranean region, is subject to intense rain events that often occur within the warm sectors of Mediterranean perturbations during the autumn season (Jacq 1994; Rivrain 1997). These perturbations are generated by upper-level troughs extending from the United Kingdom down to the Iberian Peninsula. These synoptic features lead to the advection of warm and humid air from the Mediterranean Sea toward the coastal regions. The pronounced relief (Pyrenees, Massif Central, and Alps) triggers convection and channels the low-level flows inducing low-level convergence which contributes to the release of convective instability. Various precipitating systems may result ranging from orographic systems associated with shallow convection up to mesoscale convective systems (MCS) associated to deep convection. The latter events, generally presenting a V-shape in the satellite and radar imagery, are especially productive due to their high rainfall intensities and spatial stationarity.

The following rainfall measurement devices, operated during HIRE-98, have been used in the present study (see Figs. 1 and 2):

- A network of 25 telemetering tipping-bucket rain gauges, managed by the Division Hydrologie Urbaine of the technical services in charge of the management of the Marseille combined sewer system. This network has a density of about one rain gauge per 10 km². The rain gauges have a time resolution of 6 min and a tip is equivalent to 0.2 mm of rain.
- An S-band weather radar, belonging to the Météo France ARAMIS network, located at Nîmes, 90 km
northwest of Marseille. This radar system performs three plan position indicator (PPI) scans every 5 min. The lowest elevation angle measurements (0.6°) are used to elaborate the 1 km × 1 km operational hydrological products used herein for the Marseille region.

- An X-band vertically pointing radar, installed at the Vernet site downtown. This radar system belongs to the Water Management Research Centre (WMRC), University of Bristol, United Kingdom (Cluckie et al. 2000). The height-time indicator (HTI) data have a time resolution of 4 s and a vertical resolution of 7.5 m. Table 1 provides additional information concerning the two radar systems.

The basic idea of the work is to determine a correction factor for the S-band radar that accounts for the screening of the radar beam (geometrical calculation) and for the VPR measured with the vertically pointing radar. The raw and corrected S-band radar time series are then compared to the observed rain gauge time series for various integration time steps. The drop size distribution (DSD) parameterization is known to have an important impact in terms of rain-rate estimation (e.g., Joss and Waldvogel 1990); specific $Z$–$R$ relationships derived from two DSD models fitted on Mediterranean datasets will then be considered herein to test the relative influence of the $Z$–$R$ relationship and the vertical structure of precipitation. Note that there is certainly a relationship between the DSD and the VPR (Sempere-Torres et al. 1999): however, this subject requires detailed investigations that are beyond the scope of the present paper. In section 2, some theoretical developments related to consideration of the vertical heterogeneity of the rain field in the weather radar equation are first provided. The HIRE-98 dataset used in this study is then presented in section 3. The results are given and discussed in section 4 in terms of (i) the “hydrologic visibility” (Pel larin et al. 2002) of the Nîmes radar in the region of Marseille, (ii) the radar–rain gauge assessment at the X-band radar site, and (iii) the spatial representativity of the VPR correction. Section 5 is devoted to the conclusions of this work.

### 2. Theory

The purpose of the present section is to derive a theoretical expression for the distant S-band weather radar rain-rate estimates accounting for both the vertical heterogeneity of the rain field and possible screening effects occurring between the radar and the target of interest.

Let $M_0$ be the center of the radar resolution volume and let $M$ be the center of an elementary volume containing scattering elements (hydrometeors) that contribute to the power sampled at $M_0$. Both the Cartesian and spherical coordinates, $(x, y, h)$ and $(r, \theta, \phi)$ respectively, will be used in the following derivations.

Owing to the fact that the vertical variations of the reflectivity field are larger than its horizontal variations, we may assume that the reflectivity $Z$ can be split into two orthogonal terms, with

$$Z(r, \theta, \phi) = Z(x, y, h) = Z_{\text{REF}}(x, y)z(h), \quad (1)$$

where $Z_{\text{REF}}(x, y)$ represents the reflectivity factor at a given reference level $h_{\text{REF}}$, for example, the ground level. The variable $z(h)$, called the vertical profile of reflectivity (VPR), represents the variations of the reflectivity factor as a function of altitude. The formulation (1) implies that, at the scale of the resolution volume...
of the radar situated at long range, the horizontal variations of the reflectivity are small [so that one can write: \( Z_{\text{REF}}(x, y) = Z_{\text{REF}}(x_0, y_0) \)] and the variable \( z(h) \) is assumed to be horizontally invariant. These two assumptions are useful for the following mathematical developments. They clearly do not hold for the smallest integration time steps (see, for instance, the instantaneous VPR time series presented in Fig. 4). It is hoped that their validity increases with increasing integration time steps.

A general expression of the radar backscattered power sampled at \( M_0 \) can be given by the following equation:

\[
P_r(r_0, \theta_0, \phi_0) = C_i \int \int \int \frac{L(r, \theta, \phi) W_\theta(r, \theta, \phi) Z(r, \theta, \phi)}{r^4} dV,
\]

where \( C_i \) is a constant depending on the radar parameters. According to radar measurement principles (Doviak and Zrnic 1993), the function \( W_\theta(r, \theta, \phi) \) weights the contribution of the scattering elements contained in the elementary volume \( dV \) centered at \( M \) to the returned power sampled at \( M_0 \). This function is the product of the two way power gain of radiation pattern \( G^2 \) \((\theta_0, \phi_0) \) and a radial weighting function \(| W(r_0) |^2 \) (with: \( r_0 = r - r_0, \theta_0 = \theta - \theta_0, \) and \( \phi_0 = \phi - \phi_0 \)) for which Gaussian approximations will be used herein (Delrieu et al. 1995; Pellarin et al. 2002). The function \( L(r, \theta, \phi) \) describes the power loss that may occur between the radar and the target of interest. The power loss may result from (i) the presence of partial or total masks due to the relief or anthropic targets and/or to (ii) cloud and rain attenuation effects. The latter effects will be neglected in the present study, a reasonable assumption for the S-band frequency.

Let us define the measured reflectivity by

\[
Z_m(r_0, \theta_0, \phi_0) = P_r(r_0, \theta_0, \phi_0) r_0^2/C,
\]

where \( C \) is the radar constant for a hydrometeorological target (Doviak and Zrnic 1993). It can be shown that the measured reflectivity can be expressed as follows when (1) is used to describe the vertical heterogeneity of the reflectivity field:

\[
Z_m(r_0, \theta_0, \phi_0) = Z_{\text{REF}}(x_0, y_0) z_m(r_0, \theta_0, \phi_0).
\]

The function \( z_m(r_0, \theta_0, \phi_0) \) accounts for the vertical structure of precipitation, the screening effects, and the convolution performed at the scale of the radar resolution volume. It is given by the following expression:

\[
z_m(r_0, \theta_0, \phi_0) = \left[ \frac{1}{C'} \int \int_L(r_0, \theta, \phi) f^2(\theta_0, \phi_0) z(h) \sin \theta \, d\theta \, d\phi \right]^{1/5}.
\]

where \( C' \) is the value of the integral at the right side of (5) when there is no power loss \((L(r, \theta, \phi) = 1)\) and when the reflectivity field is assumed to be homogeneous at the scale of the radar resolution volume. Note that in the practical evaluation of the function \( z_m(r_0, \theta_0, \phi_0) \), the altitude of the radar beam center is evaluated hereafter using the 4/3 Earth’s radius approximation.

Finally, a power-law \( Z-R \) relationship \((Z = aR^b)\) is usually considered for estimating the rain rate at the reference level, leading to

\[
R_{\text{REF}}(x_0, y_0) = \frac{Z_m(r_0, \theta_0, \phi_0)_{1/5}}{a_{\text{m}}(r_0, \theta_0, \phi_0)}.\]

In this expression, the function \( CF = Z_m(r_0, \theta_0, \phi_0)^{-1/5} \) appears as a correction factor which allows the rain rate at the reference level \( R_{\text{REF}}(x_0, y_0) \) to be corrected for screening and VPR effects. In the present configuration, the S-band radar provides the \( Z_m(r_0, \theta_0, \phi_0) \) values. The X-band vertically pointing radar allows the determination of the \( z(h) \) function required for the numerical evaluation of CF. The power loss function \( L(r, \theta, \phi) \) will be estimated with the procedure developed by Delrieu et al. (1995) to characterize ground clutter and screening effects using a digitized terrain model of the region of interest. The choice of the \( Z-R \) coefficients, related to the drop size distribution modeling, is discussed in section 3a. The quality of the radar-estimated rain rates will be assessed with respect to the measured rain-rate values at ground level provided by the rain gauge network.

3. Dataset

The three most important rain events of the HIRE-98 dataset (Uijlenhoet et al. 1999) occurring on 7 and 11 September and 5 October 1998 were selected for the present study. Prior to the presentation of the rain gauge, S-band radar (section 3b), and X-band VPR data (section 3c) for these specific events, considerations in terms of DSD modeling and the establishment of relations between the bulk variables of interest are first described (section 3a).

a. DSD modeling and resulting relationships between the bulk variables of interest

Due to the fact that weather radar systems having different characteristics, notably different frequencies, were used in the present study, special attention was paid to deriving and using consistent relationships between the bulk variables of interest (rain rate \( R \), reflectivity \( Z \), attenuation coefficient \( k \)) and the underlying DSD. The rain rate–scaled DSD formulation proposed by Sempere-Torres et al. (1994) and further developed by Sempere-Torres et al. (1998) was used for modeling measured DSD spectra coming from two datasets previously collected in Marseilles. The first one (DSD1 in the following) was collected with a Joss-Waldvogel disdrometer (Delrieu et al.
Table 2. Drop size distribution parameters $[N(D, R)]$ in cm$^{-3}$ and resulting $Z$-$R$ and $k$-$R$ relationships (with $Z$ in mm$^6$ m$^{-3}$, $R$ in mm h$^{-1}$, and $k$ in dB km$^{-1}$) obtained at S- and X-band using the Mie theory and Beard’s terminal velocity model (Beard 1976). The attenuation effects are supposed to be negligible at S-band.

<table>
<thead>
<tr>
<th>Name</th>
<th>DSD model</th>
<th>S-band</th>
<th>X-band</th>
</tr>
</thead>
<tbody>
<tr>
<td>DSD1</td>
<td>$N(D, R) = 0.0543R^{0.07} \exp(-39.9R^{-0.195}D)$ (Delrieu et al. 1997)</td>
<td>$Z = 290R^{0.35}$</td>
<td>$Z = 236R^{0.35}$</td>
</tr>
<tr>
<td></td>
<td></td>
<td>$k = 7.3 \times 10^{-3} R^{1.25}$</td>
<td>$k = 1.13 \times 10^{-2} R^{1.19}$</td>
</tr>
<tr>
<td>DSD2</td>
<td>$N(D, R) = 0.016R^{0.37} \exp(-30.0R^{-0.16}D)$ (Salles et al. 1999)</td>
<td>$Z = 592R^{0.35}$</td>
<td>$Z = 620R^{0.41}$</td>
</tr>
</tbody>
</table>

1997). The time resolution of the measurements was equal to 1 min and, in the DSD analysis, no attempt was made to sort the DSD spectra with respect to rain type. The DSD1 model can then be considered as representative of mixed Mediterranean precipitation. The second DSD dataset was also collected in Marseille with an optical spectropluviometer (OSP; Salles et al. 1998). A DSD model (DSD2 in the following) was specifically fitted on convective DSD spectra leading to a parameterization representative of convective Mediterranean precipitation. The negative exponential models fitted to the two datasets are listed in Table 2. The relationships between the reflectivity and the rain rate ($Z$-$R$) and between the attenuation coefficient and the rain rate ($k$-$R$) were established using Beard’s model (Beard 1976) for the terminal velocities of the raindrops and the Mie solutions for the backscattering and attenuation cross sections (Delrieu et al. 1999). The resulting power-law models are presented in Table 2. Note the slight frequency dependence (due to the Mie calculation) as well as the significant contrast between the DSD1 and DSD2 $Z$-$R$ relationships which will allow the sensitivity of our results to the DSD to be tested, at least in average.

b. Rain gauge network and S-band radar data

Figure 3 presents the ranked rain gauge amounts for the three rain events selected. Clearly, the 7 September 1998 rain event belongs to the category of intense rain events with total rain amounts ranging between 60 and 86 mm, for a total duration of 5 h, over most of the rain gauge network. The maximum rain gauge intensity is rather impressive with a value of 160 mm h$^{-1}$ during a 6-min time step. The other two storms are less intense: the total rain amounts vary between 16 and 40 mm in 3 h for 11 September 1998 and between 8 and 59 mm in 4 h 30 min for 5 October 1998. The maximum rain gauge intensity is 60 and 40 mm h$^{-1}$ in 6 min for the two storms, respectively. The 5 October 1998 storm presents a marked horizontal heterogeneity at the scale of the rain gauge network (two rain gauge amounts are greater than 40 mm) while the other two events appear more uniform in space.

The total rain amounts estimated with the S-band Nîmes radar at the rain gauge sites are also displayed in Fig. 3. In order to calibrate the radar data, Météo-France applies a spatially uniform correction factor derived from monthly radar–rain gauge statistics, according to an operational procedure described by Chêze and Helloco (1999). Since the mentioned procedure uses the Marshall–Palmer $Z$-$R$ relationship, we have first had to perform a recalibration in order to account for the $Z$-$R$ relationships proposed in this study (Table 2). This was done using the following formula: $Z_m' = \left\{ a/[(\alpha_{MP})^{1.03}] \right\} Z_m^{1.03}$, where $Z_m'$ and $Z_m$ are

Fig. 3. Ranked rain gauge amounts for the three selected rain events plotted together with the corresponding rain amounts derived from the recalibrated Nîmes radar data. Missing rain gauge values are set to $-10$ mm. The bias (mean radar value/mean rain gauge value) is equal to 0.56, 0.28, and 0.33 for the rain events of 7, 11 Sep, and 5 Oct 1998, respectively.
respectively the recalibrated and initial measured reflectivities $Z_m$ (expressed in $mm^2 m^{-3}$), $(a, b)$ the DSD1 or DSD2 $Z-R$ relationships coefficients, and $(a_{up}, b_{up})$ the Marshall–Palmer $Z-R$ relationship coefficients. The radar rain amounts displayed in Fig. 3 were then simply obtained by (i) converting the 5-min reflectivities into rain rates using the DSD1 or the DSD2 $Z-R$ relationship (note that both relationships lead to the same rain-rate estimates owing to the calibration procedure) and (ii) the accumulation of the rain amounts over the storm duration. The comparison of the radar and rain gauge amounts shows the important underestimation of the radar data even though the radar system is thought to be well calibrated electrically. The bias (ratio of the mean radar value and the mean rain gauge value) is equal to 0.56, 0.28, and 0.33 for the 7 September, 11 September, and 5 October 1998 rain events, respectively. Note that the bias for the 5 October 1998 rain event mostly results from a poor estimation of the highest rain amounts while the intermediate rain amounts are rather well estimated. Our point in the present paper is that a major part of such an observed bias can be explained by the vertical heterogeneity of precipitation. The X-band VPR data used to describe this vertical variability are presented in the next section.

c. X-band vertically pointing radar

Figure 4 presents high-resolution (4 s, 7.5 m) height-time indicators provided by the WMRC X-band vertically pointing radar during three 50-min sequences taken during the three rain events selected. The 7 September 1998 rain event is characterized by strong convective activity leading to rather chaotic patterns in the HTI images. The 11 September 1998 event is marked by the passage of several convective cells which seem to initiate at the freezing level made visible by a clear bright band at an altitude of about 3.7 km MSL. Such patterns are consistent with the “generating cell” concept proposed by Marshall (1953); the presence of high reflectivities in the lower leading part of the cells probably results from the DSD sorting which operates as rainfall. The HTIs observed on 5 October 1998 are characterized by a bright band at an altitude of about 2.5 km MSL and much more uniform and weak rain activity below this level compared to the previous cases. It is obvious from Fig. 4 that the three rain events offer a rather wide variety in terms of vertical structure, an interesting feature for the purpose of the present paper. In order to visualize the vertical region sampled by the S-band Nîmes radar in the region of Marseille, the position of the center of the beam (1560 m MSL) and the vertical extension of the 3-dB beamwidth accounting for the expected screening effects (1200–3000 m MSL approximately, see section 4a) are displayed in Fig. 4. Note that the bright band is above the Nîmes radar resolution volume for the 11 September event and within for the 5 October 1998 event. Furthermore, note, especially in the first two cases, the significant reflectivity enhancement below the Nîmes radar beam level. These two elements are certainly influential in the S-band radar–rain gauge discrepancies described in the previous paragraph.

Careful data processing, extensively described by Berne (2002; hereafter BDAV) was applied for correcting the X-band vertically pointing radar data. For this purpose, the dataset collected with a second X-band radar operated during HIRE-98 was used. This radar system, which belongs to the Laboratoire d’étude des Transferts en Hydrologie et Environnement (LTHE), was set up at the Vallon Dol site, that is, in the northern part of the City of Marseille (see Fig. 1). It was operated in RHI mode at a range of 11 km in the direction of the X-band vertically pointing radar. A first processing step was dedicated to the calibration of the two radar systems: for this purpose, the low-elevation RHI measurements of the LTHE radar were calibrated and corrected for attenuation using the mountain reference technique (MRT) proposed by Delrieu et al. (1997) and further developed by Serrar et al. (2000). Mountain returns available in the RHI measurement direction at 15 km of the LTHE radar site were used to estimate the path integrated attenuation (PIA) of the measurements required by the MRT. The radar reflectivities of the two X-band radars were then compared in the vicinity of the vertically pointing radar site, leading to a linear (in dBZ) calibration relation for the vertically pointing system. After the calibration step, the attenuation correction based on the inverse approach proposed by Vignal et al. (2003) was implemented for the two radar datasets. For the vertically pointing radar, the inverse approach was found to be equivalent to the Hitschfeld–Bordan algorithm (Hitschfeld and Bordan 1954) owing to the relatively small PIA in the vertical (maximum corrected PIA value equal to 4.3 dB). The data processing was shown to be effective (BDAV) in increasing the consistency between the HTIs of the two radar systems although some problems remain in the products obtained for each radar. The LTHE radar data will not be used hereafter owing to (i) uncorrectable screening effects in the lowest altitudes (<500 m) due to the presence of a small mountain in the vicinity of the radar site and (ii) the fact that this radar system did not work properly for the most intense part of the 7 September 1998 event. Concerning the X-band vertically pointing radar, the remaining problems are related to the unsatisfactory correction of attenuation effects in the brightband region and above: (i) $(Z, k, R)$ relationships valid for ice and/or snow hydrometeors should have been used in that region; (ii) furthermore, a threshold which was too high was implemented in the early reflectivity data archiving, thus preventing the attenuation correction at the highest altitudes. However, due to the vertical region sampled by the Nîmes radar, the following results are not thought to be greatly dependent on these remaining correction faults.

Some statistics concerning the normalized VPR function $z(h)$ were established. The average $z(h)$ values were calculated using time-averaged reflectivity profiles (ex-
Fig. 4. High space–time resolution height–time indicators (HTI) provided by the X-band vertically pointing radar for three 50-min periods taken during the three rain events selected: (top) 7 Sep, (middle) 11 Sep, and (bottom) 5 Oct 1998. The dashed and dotted lines show the center of the S-band radar beam and the limits of the 3-dB beamwidth at the X-band radar site, respectively. The X-band radar data is recalibrated and corrected for attenuation.

pressed in mm$^6$ m$^{-2}$, normalized by the time-averaged reflectivity value close to the ground (integrated over a 250-m height step) and converted to dB as a last step in order to limit the range of values. The 10%, 50%, and 90% quantiles of the statistical distribution of the $z(h)$ function are displayed in Fig. 5 for the three rain events and two integration time steps, namely 6 and 30 min. As expected, the variability of the $z(h)$ function is greater for the 7 September event and much lower for the 5 October 1998 event; it increases with altitude and is greatly reduced when the integration time step increases. Note that the average vertical gradient of the $z(h)$ function varies between 3.5 and 5 dBZ km$^{-1}$ in the first kilometers above ground level and that, when it exists, the bright
Fig. 5. The 10%, 50%, and 90% quantiles of the distribution of the normalized VPR function \( z(h) \) expressed in dB, for the three rain events selected and two integration time steps, namely, (top) 6- and (bottom) 30-min.

4. Results

As a first step (section 4a), we will concentrate on the characterization of the hydrologic visibility (Pellarin et al. 2002) of the Nîmes radar in the region of Marseille, that is both on the quantification of ground clutter and screening effects and on the numerical evaluation of the screening and VPR correction factors (CF) according to (5) and (6). Then, the results obtained at the X-band vertically pointing radar site in terms of quantification of the impact of the VPR on the S-band radar rain-rate estimates will be presented in section 4b. Finally, the results of the study devoted to the spatial representativity of the VPR will be presented in section 4c.

a. Hydrologic visibility of the Nîmes radar in the Marseille region

A numerical procedure (Delrieu et al. 1995) was used to simulate the ground clutter and screening effects of the Nîmes radar in the Marseille area. In brief, given the radar parameters (3-dB beamwidth, antenna pattern, scanning strategy, etc.) and a digitized terrain model of the region of interest, a refined geometrical calculation of the areas illuminated by the radar beam is performed by defining the intersection of the radar resolution volume and the relief. The angular and radial weighting functions, representing the radar power distribution around and along the beam axis, are accounted for in this calculation. For a given azimuth and elevation position, a mask grid is also progressively filled in range as masks affect the radar beam. The screening effects can be expressed both in terms of percent of the surface screened (geometrical mask) or percent of the radar power screened (power mask). Once these geometrical calculations are performed, the use of a wavelength-dependent model for the backscattering coefficient of the mountain surfaces allows the derivation of mountain return estimates. For comparison with weather echoes, it is convenient to express the ground clutter estimates in terms of equivalent reflectivity (dBZ). It was shown (Delrieu et al. 1995) that a good simulation of ground clutter requires con-
consideration of an angular size of the radar resolution volume of at least 2 times the 3-dB beamwidth. On the other hand, angular sizes greater than 3 times the 3-dB beamwidth (e.g., using the entire main lobe of the antenna beam pattern) provide marginal improvement. Note that a present limitation of the procedure is that it does not account for anthropic targets (urbanization, point targets) which are liable to produce strong sidelobe returns. Figure 6 presents the results obtained for the 0.6° elevation angle of the Nîmes radar which is used to elaborate the operational hydrological products of the Météo France ARAMIS network in the region of Marseille. At the Vernet site, a geometrical mask of about 35% is observed for the 5.4° angular size of the radar resolution volume (3 times the 3-dB beamwidth) considered in the simulation. Two areas of relatively strong mountains returns are simulated within the Marseille window, a result consistent with the observed radar reflectivity maps.

As a second step, the mask grids derived for each radar azimuth are used in conjunction with the normalized VPR data to estimate the rain-rate correction factors (CF) according to a numerical procedure developed by Pellarin et al. (2002). Figure 7 illustrates the results obtained for the 6-min time step with a display of the time series of the normalized VPR function (corrected for attenuation) together with the corresponding CF factors. Also plotted in this figure is the CF value calculated for the screening effects alone, that is, assuming \( z(h) = 1 \) in the hydrologic visibility calculation. This factor is equal to 1.08 for the DSD1 parameterization. Although the geometrical screening factor is relatively high (Fig. 6), this CF value indicates that the screening effect has actually a low impact in terms of rain-rate estimation. The numerical values given hereafter refer to the DSD1 parameterization since the results are very similar for both DSD parameterizations. For the 7 September 1998 event, the CF values have an average of 4.6 and a standard deviation of 4.5 for the 6-min time step. Here, CF values as great as 25 are observed for individual time steps: these peaks generally correspond to VPR profiles marked by a strong decrease with altitude. They mostly occur before or after intense precipitation cells and may therefore have a limited influence in terms of rain-rate estimation. However, significant fluctuations of the CF values (in the range of 1 to 6) occurring within the intense precipitation periods may, on the contrary, greatly impact the rain-rate estimation. Time integration greatly reduces the variability of the CF factors.
with an average value of 3.6 and a standard deviation of 2.4 for the 30-min time step. A similar behavior is observed for the 11 September 1998 event with an average CF value of 5.7 (3.1) and a standard deviation of 6.1 (0.53) for the 6-min (30 min) integration time step. A much lower temporal variability [average value of 2.6 (2.4)], standard deviation of 0.82 (0.56) for the 6 min (30 min) integration time step is obtained for the 5 October 1998 event as a result of the more uniform time structure of the VPR. An important question is whether these CF fluctuations, which are especially high for the shortest integration time step, are physically meaningful. The answer is given in the next section.

**b. Radar–rain gauge assessment at the vertically pointing radar site**

Four methods for estimating the Nîmes radar rain-rate time series are evaluated hereafter with respect to the rain gauge time series measured at the vertically pointing radar site:

- **Method 1 (M1):** the Nîmes radar reflectivities are simply converted into rain rates using the $Z$–$R$ relationship.
- **Method 2 (M2):** the Nîmes radar reflectivities are corrected by the CF factors accounting for the screening effect and the VPR determined with the calibrated X-band radar reflectivities (no attenuation correction).
- **Method 3 (M3):** identical to M2, but the CF factors...
account for the screening effect and for the VPR calibrated and corrected for attenuation.

- Method 4 (M4): identical to M3, but the CF factors are evaluated with the VPR normalized by the reflectivity at an altitude of 1000 m MSL and set to 1 at lower altitudes.

Method 1 is the basic estimation method for which the screening and VPR effects are not corrected for; remember that the Nîmes radar is thought to be correctly calibrated electrically. In addition, M2 and M3 account for the screening effect and the VPR; the two methods will allow the sensitivity of the results to the VPR to be tested. Furthermore, the relative performance of the two methods will give some indication on the effect of the quality of the VPR estimation and notably on the attenuation correction applied to the X-band vertically pointing radar data. By canceling the lower part of the VPR, M4 aims at evaluating the importance of the low-level enhancement of the reflectivity, which is actually not observable by the Nîmes radar at the considered range.

Three criteria were used for assessing the four estimation methods: (i) the bias \( B \), evaluated as the ratio of the average radar time series to the average rain gauge time series; (ii) the determination coefficient \( \rho^2 \), that is the square of the correlation coefficient, as a measure of the covariation of the two time series; and (iii) the Nash criterion (Nash and Sutcliffe 1970), defined as

\[
N = 1 - \frac{\sum_{i=1}^{n} (X_i - X^*_i)^2}{\sum_{i=1}^{n} (X_i - \bar{X})^2}, \tag{7}
\]

where \( (X_i, i = 1, n) \) is a series of reference values of a given physical process with mean \( \bar{X} \) and \( (X^*_i, i = 1, n) \) is a series of estimated values of \( X \). The Nash criterion, or efficiency, is quite popular in the hydrological sciences. It is in particular employed in the context of parameter optimization of hydrological models since it has the definite advantage of being sensitive to both the average values (like the bias) and the covariation (like the determination coefficient) of the estimated and reference series. Note that \( N = 1 \) denotes perfect agreement between the two series; \( N = 0 \) when the estimator \( X^* \) provides estimates as poor as the simple average of the reference values (the Nash criterion can also be viewed as a criterion showing the improvement of the estimator \( X^* \) over the estimator \( \bar{X} \)); a very poor agreement between the estimated and the reference values is likely when \( -\infty < N < 0 \).

Table 3 lists the values of the criteria obtained by the four estimation methods for a given integration time step (e.g., 6 min). As already mentioned in section 3b, the raw radar estimates (M1) are strongly biased \( (B = 0.31) \); the covariation of the two time series is however not too bad with a determination coefficient equal to 0.69. The Nash criterion is in that case essentially affected by the bias of the radar time series and presents a very low value close to 0. Note that the screening correction marginally improves the assessment criteria with a bias of 0.34 and a Nash coefficient of 0.11. The VPR corrections M2 and M3 appear very effective in reducing the bias (0.94 and 0.84, respectively) and provide a very significant improvement in terms of covariation of the radar and rain gauge time series (determination coefficient of 0.87; that is a gain of 18 points of explained variance for this integration time step). Although the bias is greater for M3 and the determination coefficient equal for M2 and M3, the Nash criterion indicates a significant superiority of M3 over M2. This is the result of the better repartition of the points, and notably the highest values, around the bisecting line for the M3 method (see Fig. 8) while the M2 scattergraph is characterized by underestimation (overestimation) of the rain rate values less (greater) than 50 mm h\(^{-1}\). This subtle effect illustrates the interest of using several criteria and, notably the Nash criterion, for comparing estimated and reference series. The superiority of M3 over M2 is comforting regarding

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Table 3. Assessment criteria of the radar rain-rate time series with reference to the Vernet rain gauge time series for three integration time steps and the two DSD parameterizations. The criteria are evaluated globally for the three rain events selected. In each box, the four lines correspond to (i) the bias (B) estimated as the ratio of the mean radar value and the mean rain gauge value, (ii) the square of the correlation coefficient ($\rho^2$), (iii) the Nash criterion ($N$), and (iv) the number of pairs ($n$) taken into account in the criteria evaluation. Note that a threshold of 1 mm h$^{-1}$ was considered in the rain-rate evaluation criteria. The radar rain-rate time series are issued from the recalibrated Nîmes radar data (Method 1) and from three correction methods accounting for both screening and VPR effects. The first correction method (Method 2) makes use of the VPR data provided by the vertically pointing radar simply recalibrated. The second (Method 3) uses the VPRs recalibrated and corrected for attenuation. The third one (Method 4) uses the VPRs calibrated and corrected for attenuation but normalized by the reflectivity value measured at 1000 m MSL and artificially set to 1 below this altitude. This last simulation aims at evaluating the importance of the low-level enhancement of the reflectivity which is not observable by the Nîmes radar at such a range.

<table>
<thead>
<tr>
<th>Integration time step (min)</th>
<th>DSD parameterization</th>
<th>Recalibrated radar data (Method 1)</th>
<th>VPR-corrected radar data (Method 2)</th>
<th>VPR-corrected radar data (Method 3)</th>
<th>VPR-corrected radar data (Method 4)</th>
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<tbody>
<tr>
<td>6</td>
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<td>B = 0.31</td>
<td>B = 0.94</td>
<td>B = 0.84</td>
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<td>n = 102</td>
<td>n = 102</td>
<td>n = 95</td>
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<td></td>
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The validity of the attenuation corrections applied to the X-band VPR data: the CF values, significantly too high for M2, are nicely reduced thanks to the attenuation correction which diminishes the vertical gradients of the VPR. The results of the M4 estimation method are also very interesting: compared to M1, the bias is reduced to 0.43, while the determination coefficient and the Nash criterion increase (0.74 and 0.36, respectively). However, the M4 performance is much lower than that of the M2 and M3 methods. This is a clear indication of the major influence of the low-level enhancement of the reflectivities. This result may be specific to the Mediterranean rain events which are characterized by strong vertical reflectivity gradients in the first 1000 m of altitude, probably as a result of the high humidity of the low-level air masses.

As usual, time integration has a positive impact on all the criteria, for all the methods. The relative performance of the different methods remains unchanged: the main gains are obtained on the bias criterion for the highest integration time steps. The M3 method confirms its superiority over the M2 method. A significant bias remains regardless of the time step for M3 (0.84 for 6 min; 0.80 for 30 min), an indication of remaining sources of error (calibration, DSD parameterization, . . .). The M4 method clearly fails to correctly account for the VPR effect.

c. Spatial representativity of the VPR correction

The present experimental configuration has also allowed us to empirically assess the spatial representativity of the VPR estimates derived from a point sensor like the X-band vertically pointing radar. For this purpose, the VPR time series estimated at the Vernet site were used to evaluate CF time series for each single rain gauge location in the Marseille area (except for those prone to ground clutter), which were then used to produce corrected radar rain-rate time series. The set of three criteria already used in the previous paragraph were evaluated over the available time series (three rain events) grouped in various classes of range from the X-band pointer, namely: 0 (Vernet site), [0–3 km], [3, 6 km], [6, 9 km], [9, 12 km], [12, 15 km], and [15, 18 km], with rain gauge numbers of 1, 4, 6, 4, 5, 2, and 1, respectively. The results displayed in Fig. 9 are limited to the evolution of the Nash criterion as a function of range for the M1 and M3
methods and the three integration time steps of 6, 15, and 30 min: the Nash criterion corresponding to the M1 radar estimates is pretty constant regardless of the range and the integration time step as the result of homogeneous error sources at the scale of the Marseille area. For all the integration time steps, the M3 method brings a significant improvement over the raw radar estimates at the X-band radar site. This improvement remains very limited in range, however, with typical values of 2.5, 6, and 8 km for the 6, 15, and 30 min, respectively, if one considers the range where the Nash coefficient of the corrected radar data series is greater than that of the raw radar data series. The spatial representativity of the punctual VPR estimates can then be evaluated to about 20, 100, and 200 km$^2$ for the 6-, 15-, and 30-min integration time steps, respectively.

5. Conclusions
The present study was aimed at demonstrating the major influence of the temporal variations of the vertical heterogeneity of rainfall on radar rain-rate estimates at short time steps. For this purpose, an X-band vertically pointing radar system and a network of 25 rain gauges, deployed at a range of 90 km from an operational S-band radar system, were operated during the HIRE-98 experiment. Correction factor time series for the S-band radar measurements were calculated accounting for the screen-
Fig. 9. Spatial representativity of the VPR correction. Evolution of the Nash criterion calculated between radar and rain gauge rain-rate time series for various integration time steps [(top) 6 min, (middle) 15 min, (bottom) 30 min] as a function of the distance to the X-band radar site where the VPR is measured. The dotted line corresponds to raw radar measurements (M1) and the solid line to VPR-corrected radar measurements using the X-band VPR estimates (M3). The rain-rate time series of the three rain events were considered all together and the rain gauges sites grouped in classes of distance in order to increase the robustness of the statistics.

...ing effects and the vertical heterogeneity of rainfall characterized using the X-band vertically pointing radar system. Comparison of the raw and corrected S-band radar estimates with the available rain gauge measurements for various integration time steps (6–30 min) clearly confirmed the influence of the VPR at the X-band radar site: considering about 12 h of intense Mediterranean precipitation, a continuous increase of the Nash efficiency for the corrected radar data (0.85 for the 6-min time step up to 0.93 for the 30-min time step) was observed while this criterion remains less than 0.15 for the raw radar data, regardless of the time step. These results nicely validate the “hydrological visibility” concept proposed by Pellarin et al. (2002). The bias reduction and the co-fluctuation improvement for time steps as short as 6 min clearly show that high-frequency fluctuations of the VPR explain a great part of the fluctuations in the radar–rain gauge assessment factor. The relevance of the attenuation correction applied to the X-band vertically pointing radar was also demonstrated. From a complementary viewpoint, the spatial validity of the VPR correction derived from a point sensor like the vertically pointing radar was also investigated (i) by applying the X-band–derived VPR correction factors to all the radar time series corresponding to the available rain gauge sites, and (ii) by studying radar–rain gauge assessment criteria as a function of range from the X-band profiler. The efficiency of the VPR correction remains very limited in range, although this increases with increasing time steps (approximately 2.5 and 8 km for the 6 and 30 min, respectively).

The results obtained both in the time and space domains strongly support the idea of testing the implementation of VPR retrieval algorithms at short time steps and over geometrical domains corresponding to homogeneous precipitation types. These regions could be predetermined using classification algorithms such as proposed by Steiner et al. (1995) for convective precipitations or by Sanchez-Diezma et al. (2000) for stratiform precipitations. Specific VPR corrections and Z–R relationships could hence be applied to the high time resolution reflectivity fields prior to the accumulation procedures. The simulation of the effects of the low-level reflectivity enhancement on the radar–rain gauge assessment is also an important result of the present study. Establishing reliable VPR climatologies is certainly an important challenge in order to better account for such effects which are not observable at long range from the radars.

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