해안지역 내 원거리 레이더관측자료의 보정에 관한 연구

테노리오 리카르도*, 권병혁*, 윤홍주**, 이동인***

Radar Data Correction for Long Distance Observation In Coastal Zone

Ricardo S. TENORIO, Byung-Hyuk Kwon, Hong-Joo Yoon, Dong-In Lee

요 약

해안 지역에서 해양으로부터 들어오는 중규모 정도의 강수는 단기 예보를 위하여 원격 탐사에 의한 실시간 관측이 이루어져야 한다. 그러나 위성 관측은 강수를 상세히 묘사하지 못하기 때문에 레이더를 이용한 원거리 관측이 요구된다. 본 연구에서는 레이더로 관측된 자료가 강수역을 정확하게 정량화 할 수 있도록 원거리 관측자료의 보정을 논의한다. 반사도의 연직 변화에 기인한 오차는 평균 분포나 통계적 방법으로 그리고 반사도의 감쇠 효과는 반복 편광법으로 보정될 수 있다. 이 같은 다양한 보정을 통하여 레이더로 관측된 원거리 자료를 유용화할 수 있다.

Abstract

In the coastal zone, to draw up short and medium range weather forecasts, mesoscale pluviogenic systems coming from the sea have to be observed in real time. These observations use remote sensing. However, satellite remote sensing is not sufficient to describe pluviogenic systems; reference to radar long distance observations is indispensable. This paper deals with the corrections, which must be made to long distance radar data if the rainfall field is to be both accurately and quantitatively defined. The error due to vertical variation in the reflectivity

*아라고아스대학교 기상학과
**여수대학교 해양공학과
***부경대학교 환경대기과학과
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factor can be corrected from estimation of the mean profiles or by a climatic adjustment method. Attenuation in the propagation can be corrected by an iterative polarimetric method. These various corrections permit the distance validity limits of radar data to be extended.

I. INTRODUCTION

In the coastal regions which are subject to a general circulation coming from the sea, weather forecasting presents particular difficulties since the systems of "meteorological interest", i.e. frontal depressions, come from an area, the sea, for which primary observation data, those from which forecasts are drawn up, are rare, even non-existent. This problem especially affects pluviogenic systems for two reasons. Firstly, knowledge and forecast of rainfall are one of the meteorological quantities, which are extremely useful for human activities. Furthermore, the principal characteristics of these systems are of medium scale, which is badly taken into account by forecasting models, since their space-time resolution is insufficient.

A typical example of this configuration is the coastal zone of the Brazilian "Nordeste". During the rainy season, the general circulation in this region is dominated by a southeast oceanic flow and most of the pluviogenic systems are of a frontal nature.

To overcome these difficulties, the most obvious way would be to fill in the "forecasting void" with local observations from remote sensing, either from space from a satellite platform, or from the coast by radar, or better, by combining both these types of observations. Remote sensing information is used to draw up short-range forecasts or nowcasting and is also used as a source of pluviogenic data for statistical applications.

The aim of this present article is to discuss the conditions under which radar can provide useful data for short-range local weather forecasting.

II. RADAR-SATELLITE COMPLEMENTARITY

B. Guillot's article, deals with the satellite aspects of rainfall estimation. The satellite data, which is useful for the short-range forecast of rainfall, is the infrared data provided by geostationary satellites. The observation frequency must be adapted to the velocity at which medium scale meteorological structures develop and the data must not be dependent upon a diurnal cycle. The radiometers, which are used to observe the Earth's atmosphere from space, operate in the thermal infrared, in atmospheric transparent windows, between 3.5 m and 4.2 m and 10.5 m and 12.5 m. In these frequency domains clouds are opaque, so radiometers measure the surface temperature of the cloud tops, which can be interpreted in terms of height of the top layer[1]. No direct information on the structure of the dynamic field or on the underlying rain field is obtained.

On the contrary, with microwaves, clouds are partly transparent. It is therefore possible to observe the volume distribution of precipitation enclosed in the clouds. Figure 1 illustrates these differences. It represents the distribution of the equivalent radar reflectivity factor Z_e in a vertical plane when typical convective development is present. Z_e, the equivalent radar reflectivity factor, is the quantity measured by radar. It is proportional to D^6 (Rayleigh diffusion) where D is the equivalent spherical diameter of the diffusing hydrometeor. Z_e also is dependent upon the dielectric factor $|K|^2$ of the hydrometeor. $|K|^2$ is approximately five times weaker for water than for ice. Z_e is usually expressed in $mm^6\ m^{-3}$ or in dBZ such that:

$$Z \text{ (en dBZ)} = 10 \log [Z \text{ (en } mm^6\ m^{-3})]$$

This study was carried out using the RABELAIS radar[2] from the *Aquitaine* shore, in southwest France, of 204 azimuthally. The clouds are therefore situated over the near Atlantic. They move northeastward, towards the French shores. It is a very active convective system of a frontal nature, sometimes called the mid-latitude squall line. Figure 1 only shows a part of the frontal zone. The principal convective belt is to the rear. On Figure 1, above 9 km altitude, a layer of cirrus, about 2500 m thick, can be seen, whereas below there is a "deep" convection cell of the cumulonimbus type, whose height reaches 10 km. The cirrus clouds are of the "cumulonimbo-genitus" type; they are due to the extension (or stretching) at the tropopause level of the cumulonimbus anvil of the principal convective line, due to an important high southwest flow. They are made up of ice-crystals whose drop speed is very low. The rainfall rate, which is associated with such, a cloud is around zero. The observation of the Doppler velocities (not shown) indicates that the vertical movements of the air in the cirrus layer are almost zero.

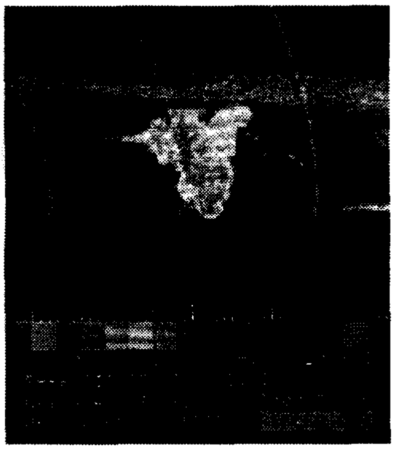


Figure 1. Distribution in the vertical plane (IHR), of the equivalent radar reflectivity factor Z_e. In the presence of a prefrontal precipitating convective cell surmounted by a cirrus layer, observed on 14 august 1987 at 0 h 56 min. of 204 azimuthally from the French Atlantic coast. Z_e is represented according to a logarithmic scale in dbz. The color intervals are separated by 5 dbz. The circles are markers of distance spaced out at 10 km intervals.

The convection cell below 10 km altitude is, on the contrary, animated by an intense updraft (greater than 20 s m^{-1}) and generates rainfall throughout its height. The maximum intensity near the surface estimated from radar reflectivity according to the relation[3] is:

$$Z = 228R^{1.62} \qquad (1)$$

where Z is expressed in mm^6 m^{-3} and R in mm h^{-1}, is of R = 10 mm h^{-1}.

In the presence of such a configuration, the satellite infrared only records the temperature field, both cold and uniform, of the upper side of the cirrus, which is characteristic of a stratiform situation, which does not generate heavy precipitation. It therefore does not take underlying convection into account.

Such differences between the infrared and microwave response can be explained by the diffusion properties of electromagnetic waves as a function of their wavelength. The key parameter in this diffusion is a, the ratio of diameter D (or its spherical equivalent) of the hydrometeor diffusing at λ, the electromagnetic wavelength of the sensor under study, such that:

$$\alpha = \frac{\pi D}{\lambda} \qquad (2)$$

where D and λ are expressed in the same unity, a is also called the radioelectric size of the diffusing particles. Figure 2 represents the variation, as a function of , of the effective section of retrodiffusion , as it results from the use of the Mie[4] equations. The curve demonstrates that diffusion by water or ice clouds on the thermal IR is approximately 10^6 to 10^{10} times greater than that of the microwaves. This is why the "condensed atmosphere" is opaque on the thermal IR. This opacity is even more important on the visible channel since the wavelength is shorter. Therefore, it is necessary to refer to microwaves to

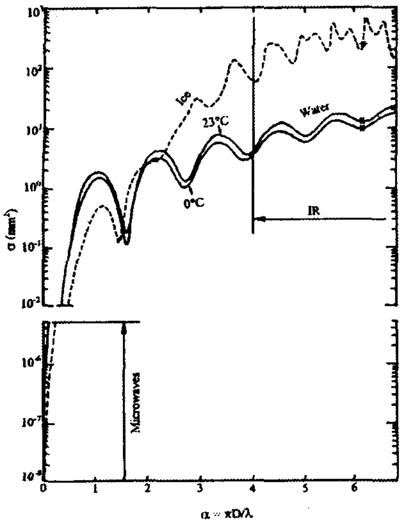


Figure 2. Efficient section of Mie diffusion as a function of the radioelectric size of hydrometeors (α) with identification of microwave diffusion domains of the radar and the thermal IR by clouds for equivalent diameters of particles between 20μm and 200μm.

visualize rainfall. If suitable microwave spatial instrumentation is not available, land radars are used.

III. TROPICAL PLUVIOGENIC SYSTEMS

Figure 1 does not illustrate the general case. It corresponds to an intermediary stage in an ordinary convective development selected for its didactic interest. Generally, the tropical convective clouds, which generate rainfall, following a rapid upward growth phase, come up against the tropopause above which the atmosphere is statically stable. Not being able to continue its upward movement, the ascending current diverges horizontally and the cloud spreads out under the tropopause. Where the updraft "collides with" the tropopause, one can observe overshooting of the cloud top in relation to the mean top level. This overshooting is due to the kinetic energy acquired at lower levels drawing up ascending air beyond its static equilibrium altitude. This overshooting is only temporary.

Figure 3 represents a typical tropical squall line which was observed in West Africa in the Sudanese-Sahelian zone. This is the preferential type of intense pluviogenic tropical convection organization. One of the essential characteristics of squall line organization is that the convective cells form a compact line at the front of the system whereas the rear is occupied by a vast stratiform zone. The convection line is sometimes associated with the overshooting cited above. However, in general, the system top appears on IR images as a vast region with uniformly cold temperature. In fact, convection can exist without overshooting being present or detectable and analysis of radio-metric microwave data (on polar orbiting satellites such as the one which bears the SSM/I radiometer) demonstrates that heavy rains are not always where one expects them to be[5].

Squall lines as well as mesoscale convective clusters or isolated thunderstorms are made up of a group of convective cells. These cells can in general be individualized or identified in the radar reflectivity field by relative maxima. The activity cycles of individual cells making up a group are clearly not synchronous. On the contrary, they are out of phase, so that, within a group at the mature stage, cells reaching the end of their life are replaced by new cells. This regeneration mechanism participates in a non-negligible manner in the displacement of the group. The life span of an individual cell is usually in the order of 20 minutes, whereas the life span of the group is from a few to more than 24 hours.

In conclusion, if the satellite IR information shows indisputable potentials to estimate rainfall from space and to globally monitor the situation, it does not permit, due to the present state of interpretation algorithms, the situation to be observed with sufficient accuracy to ensure a short-range forecast. The land radar would appear to be indispensable to complete these insufficiencies to quantitatively describe, with appropriate time and space resolution, the development of the local precipitation field associated with the great tropical systems.

IV. DIFFICULTIES LINKED TO RADAR LONG DISTANCE OBSERVATION

A "standard performance" land meteorological radar is capable of (quantitatively) detecting rain systems associated with dense convection over considerable distances, up to 300 km to 400 km in flat areas. Consider, for example, the characteristics of Thomson-CSF (France) RODIN radars, or Gematronic METEOR 300 AC (Germany), or EEC WR 100 (U.S.A.), which are fairly alike Table 1.

Table 1. General characteristics of the C-band radar.

Parameters	Values
Wavelength (λ)	5.5 cm (C-band)
Power peak (P^c)	250 kW
Pulse width (τ)	2 μs
Return frequency (non-Doppler mode)	330 Hz
Beam width at 3 dB	1.2°

The detection threshold of such radars is approximately 1 mm h^{-1} at a distance of 200 km. However, it must be highlighted that the conditions under which a land radar scans the atmosphere change according to the distance between the radar and the "target". As the radar-target distance increases, the land radar beam moves away from the ground. This is due to the rotundity of the earth but also to the fact that in order to avoid (or lessen) ground echoes and total or partial occultations by terrestrial obstacles, the radar beam explore the environment with a non-zero site angle (or elevation above the horizon). For instance, in Figure 3a, the site angle is 0.9. But, the precipitation reflectivity, for the same intensity of rainfall at the ground level, varies with altitude. Figure 3b clearly shows the variation in the reflectivity profile with height above ground level. On average, reflectivity decreases with height in the rain layer, it presents a relative maximum at the fusion layer level (i.e. over a height of a few hundred meters below the 0C isotherm), then once more decreases with height, above the fusion layer. When the sought information is the rainfall intensity at the ground, that is R_s, it is necessary to correct for the effects of this reflectivity vertical profile (RVP).

A second cause of error in the radar measurement is the attenuation of the signal by gases, clouds and precipitation. This attenuation is dependent upon the wavelength of the radar. It increases when decreases. It is almost negligible on the S-band (λ =10 cm), but can be prohibitive in the X-band (λ=3.2 cm). This is why most land meteorological radars operate in the C-band (λ=5.5 cm) since the attenuation is moderate here, except in rainfall when it is extremely heavy. Moreover, up to now, the quantitative use of radars has been limited to moderate radial distances

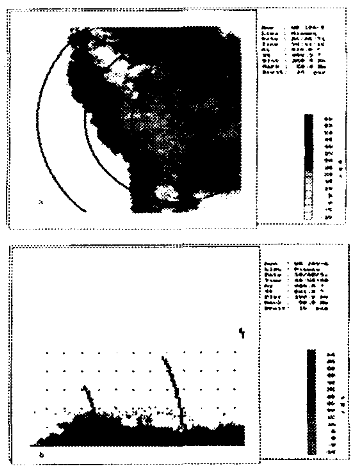


Figure 3. Distribution of the equivalent radar reflectivity Z_e. Associated with a Sudanese-Sahelian squall line observed with the Niamey Niger C-band radar and the Sanaga acquisition system (original image in color converted into gray density variations), a) PPI, b) RHI.

(from 100 km to 150 km). However, it is now possible to greatly increase these limits when the two problems cited above can be mastered.

V. CORRECTION FOR THE EFFECTS OF THE REFLECTIVITY VERTICAL PROFILE (RVP)

Two techniques have been put forward to correct for the effects of the reflectivity vertical profile. One is based on estimating the correction profile, the other is a statistical adjustment.

5.1. Estimation of a correction profile

This method is based on the hypothesis that the observed precipitation field is stationary [6, 7]. It is assumed that the ratio between the values of the mean equivalent radar reflectivity factor to altitude z, that is $Z_e(z)$, and to the ground, that is $Z_e(z_o)$, does not vary during the time taken to estimate RVP nor during its application to correct measurements. Provided that this hypothesis is acceptable, one may use radar reflectivity data at various distances and at various altitudes where there is detectable precipitation, to define a reference profile, that is f(z) this profile. To reduce the effects of space-time variability of precipitation fields on the f(z) profile, one can consider, at each z level, the mean of the available values at this level. Therefore, one averages the observed values in the annular domain situated between distances r and r + Ar, that is to say between altitudes z and z + Az. Also, one averages during time (sliding mean). The value of the equivalent radar reflectivity factor, corrected at the surface level, is simply obtained by the expression (by definition):

$$Z_e(z_0) = f(z)Z_e(z) \quad \text{......} (3)$$

If $Z_e(z_o)$ is known, one can calculate $R(Z_o)$ by an adapted Z_e-R relation.

This method has not yet been sufficiently tested for its application limits to be precisely known. However, it is clear that it is better adapted to the frontal systems of mid-latitudes, whose horizontal extension is important and whose structure is relatively homogeneous, than to the tropical convective systems. When there are isolated convective structures, the due sample to estimate f(z) is not always available. In the case of squall lines, the profile is not the same in the line of intense convection or in the stratiform zone.

5.2. Climatic adjustment

The correction method entitled "climatic adjustment" is based on the fact that the probability distribution of the precipitation intensity P(R) in a given region which is climatically homogeneous, is a well defined function which can be determined with a relatively limited observation sample and which, inversely, can be used to represent an observation sample also limited. Figure 4 shows, as an example, the probability distribution of rainfall intensity, that is P(R), for the southern Niger climatic area in West Africa (Programme EPSAT) deducted from 1988 and 1989 disdrometer measurements[8]. In the case shown in Figure 4, P(R) was calculated from measurements based on a very small surface, that of the sensor (50 cm^2) but with a very long observation time (6912 distributions of drops integrated over a time interval of 1 minute). P(R) can also be defined from a single radar panoramic scan if the observed precipitation field is sufficiently large and if the appropriate Z-R relation is known.

It can be demonstrated that in order to define the Z-R relation with satisfactory stability, it is only necessary to consider a sample corresponding to a quantity of total rainfall in the order of 600 mm[9]. Therefore, for instance, if, in a radar panoramic scan, 10,000 pixels of 1 km^2 with a precipitation signal in each of them are observed, that is a precipitation area of only 100 x 100 km^2, P(Z) can be calculated.

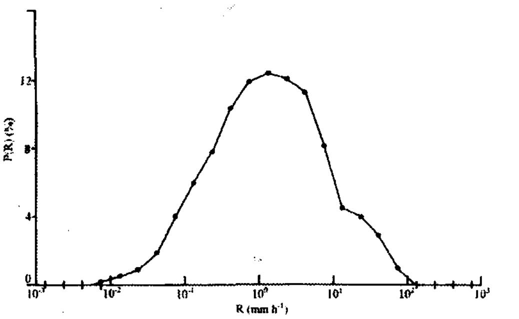


Figure 4. Probability distribution of the rainfall intensity logarithm at Niamey in Niger, from disdrometer measurements made in 1988 and 1989. 6912 distributions, measured over a time interval of 1 minute were taken into account (according to [9]).

If the true ground distribution $P(R, z_o)$ is known, having defined it with the help of non-biased ground measurements (for example with a disdrometer, or with a calibrated radar observing rainfall just above ground level). The climatic adjustment method[10] simply consists of noting that the P(Z, z) distribution observed at altitude z must correspond to the $P(R, z_o)$ distribution and that the correction can be obtained by comparing both distributions. P(Z, z) is defined by considering the observed sample in the annular domain situated between altitudes z and z + z (between the radial distances r and r + r). To obtain the coefficients a and b belonging to the relation Z-R permitting Z(z) to be converted into $R(z_o)$, it can be noted that;

$$R(z_o) = \left(\frac{1}{a} Z(z)\right)^{1/b} \quad \text{...} (4)$$

it is only necessary to regress the values of log $[R_i(zo)]$ and log $[Z_i(z)]$ corresponding to equal probability thresholds i. These values are calculated by integrating the probability distributions $R(z_o)$ and Z(z) between 0 and R_i or Z_i respectively, in such a manner that the integrals are equal at threshold i. However, these coefficients can be obtained more simply. In fact, it can be observed that in nature, P(R) is a log-normal distribution[8, 11]. Consequently, if it is assumed that Z and R are linked by a relation in the form Z = aR^b, P(Z) is also a log-normal distribution. If μ and σ denote the mean and the standard deviation of the distribution respectively, one finds [12], taking into account that $\sigma_R^2 \cong 1.8$ [8], then:

$$\ln a = \mu Z - 0.75 \mu R \sigma Z \quad \text{.......................................} (5)$$

$$b = 0.75 \sigma Z \quad \text{...} (6)$$

The Z-R relations so obtained, for each altitude layer, permits the RVP effects to be corrected for.

An even simpler method, derived from the same concepts, consists of converting Z(z) into R(z) with the help of an approximate Z-R relation, from the literature, then to define for each z level, a correction factor from the ratio of the climatic mean precipitation intensities, that is $\mu_o(z_o)$, measured with the radar, that is $\mu(z)$.

These climatic adjustment methods have the advantage of also taking into account errors due to instrumentation (radar not calibrated, attenuation by the radome, etc.).

VI. CORRECTION OF THE ATTENUATION

6.1. Causes of attenuation

The causes of attenuation associated with "radar propagation" in the atmosphere are well known. They are the atmospheric gases (O_2 and H_2O), clouds and precipitation.

a) Gas

When precipitation is observed at moderate distances, attenuation by gases is often negligible. For long distance observation, attenuation must be taken into account. For example, in the C-band, for an observation distance of 200 km, attenuation by molecular oxygen near the surface is approximately 2.8 dB, attenuation by water vapor for a content of 25 g m^{-3}, which is a reasonable value above a tropical ocean, reaches 2.5 dB, that is a total attenuation of 5.3 dB. Taking into account the angle of the beam path which conveys it, with distance, in the atmospheric levels of lesser density, this value can be reduced. However, it is still preferable to correct it. An accurate correction would require the H_2O distribution to be known along the beam path. If this information is not available, an approximate correction can be obtained by considering propagation through a Standard Atmosphere[13].

b) Clouds

Only liquid water clouds significantly intervene in the attenuation of microwaves. Usually, attenuation by ice-clouds is negligible, due on the one hand to the fact that for the same quantity of water, their attenuation coefficient is 10 to 100 times less than that of liquid clouds, and on the other hand, to the fact that their water content is low. For instance, the signal from a target 200 km away, observed through an ice-cloud with a water content equal to 0.5 g m^{-3}, using a C-band radar, would be subject to an attenuation of less than 0.2 dB.

However, water clouds cause a real problem since they greatly attenuate but have very low reflectivity. In the C-band, for instance, the attenuation coefficient per water cloud at 10C equals:

$$A_c \cong 2\times10^{-2} M \quad \text{.........} (7)$$

where A_c is in dB km^{-1} for a path and M, the cloud water content, is in g m^{-3}.

Therefore, observation of a target situated at 200 km away through a cloud having a mean water content of M = 1 g m^{-3} is 8 dB, which is considerable.

The radar reflectivity factor of water clouds as a function of M is given by[14] :

$$Z \cong 6.8\times10^{-2} M^{1.9} \quad \text{.........} (8)$$

where Z is expressed in mm^6 m^{-3} and M in g m^{-3}. When M = 1 g m^{-3} the value of the reflectivity factor is only Z = -11 dBZ. Such weak reflectivity is not detectable on most of the radars used to observe precipitation. Moreover, cloudy air also usually contains precipitations. But rainfall, since the radar operates in the Rayleigh diffusion domain ($Z \alpha D^6$), is a major factor contributing to reflectivity. For instance, for comparison with 8, the rainfall reflectivity factor whose granulometric distribution is that of [15] is expressed as:

$$Z \cong 2.4\times10^{4} M^{1.8} \quad \text{.........} (9)$$

with the same units as (8). For the same water content (M = 1 g m^{-3} is equivalent to R = 19.9 mm h^{-1}), the reflectivity factor of rainfall is 55 dB higher that of the water cloud.

How can the effects of a cause which is not detectable be corrected? In section 6.3 it will be seen that there is an instrumental solution when one has a polarimetric radar. In the case, which is the most current, when observations are made with a traditional radar, at a single wavelength, working in reflectivity, in a single polarization plane, no general solution has been proposed. Current rainfall estimation methods using radar do not take this cause of attenuation into account. In fact the methods in question assume that, in the observation domain, the beam remains below the cloud base and that attenuation by the cloud does not occur. This hypothesis is generally acceptable for applications over very short distances, when the condensation level is sufficiently high. For instance, for a 1 beam width emitted at sea level, whose lower edge is strictly horizontal, the upper edge will reach an altitude of 3000 m at 120 km away. When the condensation level is low and there is long distance observation, attenuation by the clouds is significant.

c) Precipitation

Attenuation by precipitations is a function of temperature, wavelength and of the thermodynamic phase of the precipitation. Solid, small size hydrometeors (ice-crystals, snow-flakes) weakly attenuate. Hail greatly attenuates. Since liquid water is a dielectric with important loss, attenuation by rainfall is important and all the more so when the wavelength is shorter. The general form of the attenuation coefficient by rainfall is:

$$A_p = kR\gamma \quad \text{(10)}$$

where k and γ are coefficients dependent upon wavelength and temperature.

Apart from in the S-band, for rainfall fields of weak intensity, attenuation by rain must always be corrected for.

6.2. Iterative correction of attenuation by moderately intense precipitation

The relation between the attenuated radar reflectivity factor $Z_a(r)$ measured in rain, at a distance r, and the coefficient of attenuation by rain A_p for a radar observation using a single polarization (usually horizontal linear), is expressed as:

$$Z_a(r) = Z_0(r) - 2\Delta r \sum_{i=1}^{n-1} A_{p,i} \quad \text{(11)}$$

where $Z_0(r)$ is the non-attenuated reflectivity factor, Z is expressed in dBZ, A_p is in dB per length unit, Δr is the interval between the equidistant sampling gates and i is the gate order number where $r = n\Delta r$. By combining a Z-R relation with (10), it is possible to express A_p as a function of Z, viz:

$$A_p = ka^{-\gamma/b} Z^{\gamma/b} \quad \text{(12)}$$

By carrying (12) through into (11), one obtains a relation thanks to which attenuation by precipitations can be corrected step by step by an iterative procedure[16]. In fact, the measurement of Z_a in a gate i permits the estimation of Ap to be calculated in this gate and therefore to estimate the attenuation which affects the following gate. However, since the value of Z used to estimate A_p is below the real value (since it is attenuated), a few iterations are necessary to reach a stable and optimum result.

This method only corrects for attenuation by rainfall, it ignores attenuation by gases and in particular, by clouds. In addition, it requires a carefully calibrated radar. [16] underlined the fact that this method gave good results in the presence of moderate attenuation, that is to say attenuation which did not exceed a few decibels for the gates affected by the strongest attenuation (i.e. the gates the farthest away from the radar). When there is important attenuation, calibration errors degrade the results in unacceptable proportions; in other words, the method becomes unstable and the results diverge.

6.3. Global correction of attenuation by differential polarimetric measurement

A global method permitting the attenuation effects of gases, clouds and rainfall to be corrected for on radar measurements of rainfall intensity, has recently been proposed [17]. In addition, this method is independent of radar calibration. On the other hand, it requires the use of a radar with polarization diversity and measurement is only possible in the presence of a measurable attenuation. The latter condition excludes the S-band but enables work to be carried out at attenuated wavelengths, especially in the K_a, X and C bands, that is to say with lighter radars (and therefore less expensive than S-band radars).

This method uses radar measurement of differential reflectivity Z_{DR} which is defined by:

$$Z_{DR} = Z_H - Z_V \quad \text{(13)}$$

with Z in dBZ. The indexes H and V denote the horizontal and vertical linear polarisations respectively. It is based on the fact that, at attenuated wavelengths, one can express the differential reflectivity measured at distance r, noted $Z_{DRa}(r)$, as being equal to the difference of a term in the form $Z_{DRs}(r)$ and a term $A_{DP}(r)$ representing the differential attenuation on the path(0, r), that is:

$$Z_{DRa}(r) = Z_{DRs}(r) - 2A_{DP}(r) \quad \text{.........................} (14)$$

Z_{DRs} is the differential reflectivity which would be measured in the absence of attenuation; it therefore only depends on the form of the drops. A_{DP} is defined by:

$$A_{DP}(r) = \Delta r \sum_{i=1}^{n-1} (A_{iH} - A_{iV}) \quad \text{.............................} (15)$$

where A_{IH} and A_{IV} are the attenuation coefficients in horizontal and vertical linear polarisation respectively.

The measurement of $Z_{DRa}(r)$ with the radar can be interpreted as a measurement of $A_{DP}(r)$ if the term in the form $Z_{DRs}(r)$ is calculated from a model. But A_{DP}, taking into account (10), can be expressed as:

$$A_{DP}(r) = \Delta r (k_H - k_V) \sum_{i=1}^{n-1} R_i^{\gamma} \quad \text{..........................} (16)$$

Therefore, A_{DP} is a precise and non-biased measurement of the rain intensity integral between 0 and r. It can be seen that knowledge of this integral permits, by an iterative procedure, at high attenuation values, to overcome the problem of instability of the correction algorithm described in section 6.2.

VII. CONCLUSION

In this paper, the conditions under which radar data at long distance can be corrected in order to restore the value fields of surface precipitation rates have been dealt with. This question has only been tackled fairly recently, that is why it has not yet obtained duly validated and universally acceptable responses; it is therefore subject to evolution.

Two methods have been proposed to correct errors due to the vertical reflectivity profile: the use of a correction function deducted from the observation of the mean vertical distribution of the reflectivity and an adjusting method for parameters of the Z-R relation from the comparison of the measured probability distribution of reflectivity with the "climatic" distribution of the intensity probability of rainfall. Since each of these methods has its merits, it is suggested that they be used together through a composite correction term.

This term can be obtained by the arithmetic mean of corrective factors resulting from the application of both methods. In the case where ground measurement means are available (for example a few rain-gauges in a fraction of the observed area), it is possible to compare R values estimated after correction for each of the two methods with ground truth, then to modify the weighting allocated to each corrective term in the composite mean to take the quality of the results into account.

As for attenuation, it has been seen that three factors have to be taken into account: gases, clouds and rainfall. Moreover, the intensity of this attenuation, and therefore the need for correction, increases with frequency. Apart from in the S-band for moderate rainfall, correction is indispensable. Contrarily to a prevailing idea, attenuation by clouds is important, even in the C-band, even when they are not detected by radars due to their low level of reflectivity. There exists an iterative procedure enabling the attenuation by rainfall to be corrected, step by step, which gives good results when the attenuation to be corrected remains below a few decibels. This procedure is unstable, and therefore cannot be used when there is high attenuation. A recently proposed polarimetric method enables, from the interpretation of differential reflectivity in terms of differential attenuation, to deduce the corrected precipitating rates of attenuation by rainfall, clouds and gases. In addition, this method is independent of radar calibration.

It can therefore be concluded that means are available to qualitatively correct radar data so that they can be used over large distances.

REFERENCES

[1] GUILLOT B., "Satellite et precipitations : contraintes techniques et physiques. Analyse de quelques methodes. Problemes de recherche et de validation", Veille Clim. Sat., 55,27-58, 1995.

[2] SAUVAGEOT H., "Radar Meteorology", Artech House, 1992, 366 p.

[3] SAUVAGEOT H., LACAUX J.P., "The shape of averaged drop size distributions", J. Atmos. Sci., 52,1070-1083, 1995.

[4] MIE G., "Beitrage zur optik truber Medien, speziell kolloidaler metallosungen", Annals of Physics, 25,377-445, 1908.

[5] JOBARD I., AND DESBOIS M., "A method combining infrared and microwave radiances for the estimation of the tropical rain", Preprint. Sixth AMS Conf. on Satellite Meteorology and Oceanography, Atlanta, 5-10, 1992.

[6] SMITH C., " The reduction of errors caused by bright bands in quantitative rainfall measurements made using radar", J. Atmos. Oceanic Technol., 3,129-141, 1986.

[7] ANDRIEU H., CREUTINJ.D., "Identification of vertical profiles of radar reflectivity for hydrological applications using an inverse method. Part I: formulation", J. Appl. Meteor., 34,225-239, 1995.

[8] SAUVAGEOT H., "The probability density function of rain rate and the estimation of rainfall by area integrals", J. Appl. Meteor., 33, 1255- 1262, 1994.

[9] ROSENFELD D., WOLF D. B., AMITAI E., "The window probability matching method for rainfall measurements with radar", J. Appl. Meteor., 33, 682-693, 1994.

[10] CALHEIROS R. V., ZAWADSKI I., "Reflectivity-rain rate relationships for radar hydrology in Brazil", J. Clim. and Appl. Meteor., 26, 118-132, 1987.

[11] ATLAS D., ROSENFELDD., SHORT D.A., "The estimation of convective rainfall by area integrals, 1-The theoretical and empirical basis", J. Geophys. Res.,95, (D3), 2153-2160, 1990.

[12] ATLAS D., ROSENFELD D., WOLFF D.B., "Climatologically tuned reflectivity-rain rate relations and links to area-time integrals", J. Appl. Meteor., 29,1120-1135, 1990.

[13] COLEA E., COURT A., KANTORA.J.,"Model Atmospheres. In Handbook of Geophysics and Space Environment", S.L. Valley ed. Office of Aerospace Res., USAF, Cambridge Res. Labs., Chapter2, 1965.

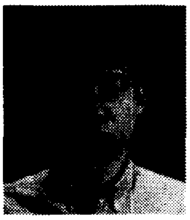
[14] SAUVAGEOT H., OMAR J., "Radar reflectivity of cumulus clouds", J. Atmos. and Oceanic Technol., 4,264-272, 1987.

[15] MARSHALL J. S., PALMER W. M. K, "The distribution of raindrops with size", J. Meteor, 5,165-166, 1948.

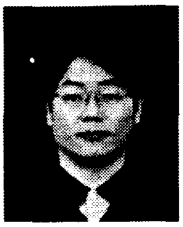
[16] HILDEBRANDP H., " Iterative correction for attenuation of 5cm radar in rain", J. Appl. Meteor., 17,508-514, 1978.

[17] SAUVAGEOT H., " Polarimetric radar at attenuated wavelengths as a hydrological sensor", Journal of Atmospheric and Oceanic Technology: Vol. 13, No. 3, pp. 630637, 1996.

Ricardo Sarmento TENÓRIO
(히까르도 사르멘또 떼노리오)
1993. 8 - 1996. 9 프랑스 Paul Sabatier University (뽈사바티에 대학교) 원격 기상학 박사
1987. 3 - 1989. 8 영국 University of Reaging (리딩 대학교) 기상학 석사
1981. 3 - 1985. 12 브라질 Federal University of Alagoas (알라고아스 주립대학교) 농업기상학 학사
현재 브라질 알라고아스 주립대학교 기상학과 조교수
기상학과 대학원 학과장
브라질 북동부 지역 레이더와 위성을 이용한 강수 연구의 책임자



권 병 혁(Byung-Hyuk Kwon)
1982~1986 연세대학교 기상학
1993~1994 그르노블대학원 지구유체역학
1998~1999 경북대학교 Post-Doc. 연구
2000~2001 알라고아스 대학교 방문교수



이 동 인(Dong-In Lee)
1980. 2부산수산대학교 환경공학과 - 공학사
1984. 2부산수산대학교 환경공학과 (대기오염) - 공학석사
1989. 3일본 북해도대학 대학원 - 이학박사
1983.10 - 1985. 3부산수산대학교 환경공학과 조교
1989. 4 - 1990. 2부산수산대학교 대기과학과 조교
1990. 3 - 1992. 2부경대학교 대기과학과 전임강사
1992. 3 - 1996. 3부경대학교 대기과학과 조 교수
1996. 4 - 현 재부경대학교 환경대기과학과 부 교수
1999. 2 - 2000. 2 미국 North Dakota 대학 교환교수
1997. 7 - 1997. 8 중국 천진 남개대학 파견
1996. 7 - 1996. 8 프랑스 Center of Research Atmosphere 파견 연구
관심분야 : 레이더 기상학, 대기열역학, 환경대기분석

윤 홍 주(Hong-Joo Yoon)
1979. 3 - 1983. 2 釜慶大學校 海洋工學科 卒業 (工學士)
1983. 3 - 1985. 2 釜慶大學校 大學院 海洋工學科 卒業 (工學碩士)
1991. 10 - 1992. 9 프랑스 Paris VI대학 博士準備課程 (海洋・大氣力學 專攻)
1992. 10 - 1993. 9 프랑스 Paris VI대학 博士準備課程(衛星遠隔探査工學, Radiometer專攻)
1993. 10 - 1997. 5 프랑스 Grenoblele I 大學 博士課程 (衛星遠隔探査工學博士 : 衛星遠隔探査工學, Altimeter專攻)
1983. 10 - 1985. 3 釜慶大學校 應用地質學科 助教
1988. 3 - 1991. 9 嶺南大學校 海洋科學研究所 常任研究員
1991. 10 - 1993. 10 嶺南大學校 海洋科學研究所 特別研究員
1993. 11 - 1997. 5
1. 프랑스 國立科學研究센타(CNRS) Grenoble 水理力學研究所 研究員
2. 美航空宇宙局(NASA) 및 佛航空宇宙局(CNES) 人工衛星 science work team 潮汐 및 海水位分野 專門研究員
1997. 8 - 1997. 12 釜慶大學校 海洋科學共同研究所 원격탐사실 委囑研究員
1997. 9 - 1997. 12 釜慶大學校 및 嶺南大學校 시간강사
1998. 1 - 1999. 8 氣象廳 기상연구소 원격탐사연구실 기상연구관
1999. 9 - 현재 麗水大學校 해양시스템학부 해양공학과 전임강사