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Structure and Evolution of Intense Tropical Cyclone Dina near La Réunion on 22 January 2002: GB-EVTD Analysis of Single Doppler Radar Observations

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ABSTRACT

Doppler radar observations of Tropical Cyclone Dina, as its eye passed at less than 100 km from the northern coast of La Réunion Island (21°S, 55.5°E) on 22 January 2002, are analyzed using the Ground-Based Extended Velocity Track Display (GB-EVTD) technique. This method is an extension of GB-VTD and it allows one to determine the full set of wavenumber-0 and -1 components of the tangential and radial winds in a tropical cyclone from a series of observations with a ground-based Doppler radar.

The results obtained for Dina reveal the presence of strong swirling winds ($>65 \text{ m s}^{-1}$) at 40–60-km radii from the storm center and below 3-km altitude. The observed changes in the location and intensity of the maximum winds, as well as the veering propagation of Dina, are shown to result probably from interaction between cyclonic winds and high topography of the island.

1. Introduction

Tropical cyclones (TCs) are threatening meteorological phenomena for islands and coastal regions in the Tropics. These perturbations are “warm core” vortices where the strongest swirling winds occur in the lowest levels at some distance from the storm circulation center. In addition to wind damage, heavy rain causes floodings, especially in mountainous regions, while high ocean tide and strong waves sweep the shores. Dedicated observations with instrumented aircraft, dropwindsondes, radar, satellites, as well as numerical models at various spatial resolution have provided valuable information on TCs. However, apart from geostationary satellite images, real-time data on TCs are relatively scarce and, even when the storm center is at relatively close distance ($\leq 100 \text{ km}$) from the threatened area, it is very difficult

to estimate the wind intensity and its spatial distribution around the storm.

Ground-based Doppler radars are essential tools to observe mesoscale precipitating systems and, in addition to providing information on rain intensity, they can be of great help to estimate the wind structure of TCs. Lee et al. (1999, hereafter LJCD), Lee and Marks (2000) and Lee et al. (2000) have shown that, through the Ground-Based Velocity Track Display (GB-VTD) technique, it is possible to deduce a plausible and physically consistent three-dimensional primary circulation of a landfalling TC using a single ground-based Doppler radar. However, the GB-VTD-derived wind description is not complete since only the symmetric part of the radial wind component can be obtained. Nevertheless, its asymmetric part is important to identify the contribution of the horizontal mean flow in the wind field. In this paper, we show that the ground-based version of the extended VTD (EVTD: Roux and Marks 1996) method—GB-EVTD—can alleviate this limitation.

The case study presented here concerns Doppler radar observations of intense Tropical Cyclone Dina near La

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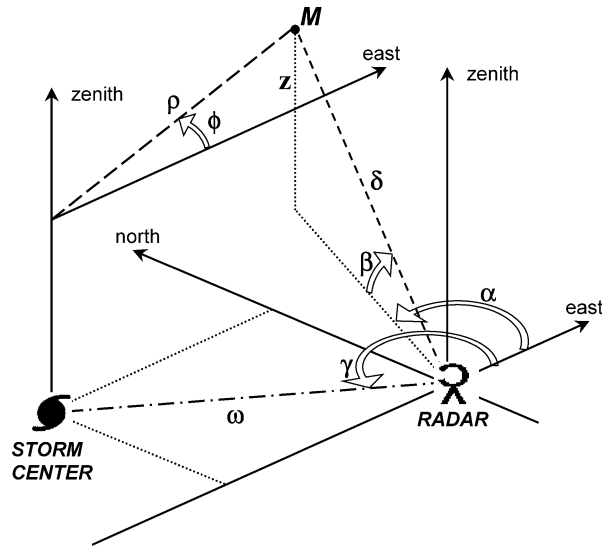


FIG. 1. Geometry of the GB-EVTD analysis. The origin of the coordinate system is at the radar location. The storm center is defined by the distance ω from the radar and the azimuth γ . Each considered point M is determined by two sets of coordinates: distance δ , azimuth α , and elevation β with respect to the radar and horizontal distance ρ , azimuth ϕ , and altitude z with respect to the storm center.

Réunion Island in the southwestern Indian Ocean on 22 January 2002. Section 2 summarizes the principles of GB-EVTD, section 3 gives a brief summary of Dina's evolution and propagation from 16 to 26 January 2002, section 4 presents the main results obtained from the GB-EVTD analysis of the Doppler data collected between 0952 (all times UTC, local time is UTC+4) and 1422, and some perspectives are discussed in section 5.

2. The GB-EVTD analysis

The VTD and EVTD analyses (initially developed for airborne Doppler observations of TCs: Lee et al. 1994; Roux and Marks 1996) consider a decomposition of the horizontal wind in the inner core region of TCs into circular harmonics supposing that (i) the air particles follow close streamlines around the storm center and (ii) the lowest harmonics of the wind components are the most energetic ones. Within a ring of given width $\Delta\rho$ and depth Δz at a radial distance ρ from the storm center and an altitude z above mean sea level (MSL), tangential V_T and radial V_R wind components can be written as

$$\begin{aligned} V_T &= T_0 + T_{1c} \cos\phi + T_{1s} \sin\phi + \varepsilon_T(n\phi, n > 1), \\ V_R &= R_0 + R_{1c} \cos\phi + R_{1s} \sin\phi + \varepsilon_R(n\phi, n > 1), \end{aligned} \quad (1)$$

where ϕ is the azimuth relative to the storm circulation center (0 is eastward, ϕ increases counterclockwise; see Fig. 1); ε_T and ε_R denote assumed minor contributions from higher wavenumbers.

Once corrected for elevation β of the radar beam, hydrometeor fall speed V_p , which can be estimated from the radar reflectivity values [we use the same relations as Gamache et al. (1993)], and storm motion (westerly U_0 and southerly V_0 , determined from the successive positions of the storm center; see below), the Doppler radial velocity V_{DOP} measured in a TC with a ground-based radar, once corrected for storm motion and hydrometeor fall speed, can be written as V_{DOP}^*

$$\begin{aligned} V_{\text{DOP}}^* &= \frac{1}{\cos\beta} (V_{\text{DOP}} - V_p \sin\beta) - U_0 \cos\alpha - V_0 \sin\alpha \\ &= V_T \sin(\phi - \alpha) + V_R \cos(\phi - \alpha) \\ &\approx T_0 \sin(\phi - \alpha) + T_{1c} \cos\phi \sin(\phi - \alpha) \\ &\quad + T_{1s} \sin\phi \sin(\phi - \alpha) \\ &\quad + R_0 \cos(\phi - \alpha) + R_{1c} \cos\phi \cos(\phi - \alpha) \\ &\quad + R_{1s} \sin\phi \cos(\phi - \alpha), \end{aligned} \quad (2)$$

where α is the radar-relative azimuth (0 is eastward, α increases counterclockwise) and β the elevation with respect to the horizontal (Fig. 1). In the following, V_{DOP}^* will be simply referred to as V_{DOP} , and wavenumbers higher than 1 are ignored.

Six unknown values (T_j , R_j , $j = 1, 3$; hereafter referred to as TR_m , $m = 1, 6$) have to be calculated within each ring of radius ρ and altitude z . This situation is similar to that encountered in the VTD analysis except the "radar relative" azimuth α is variable for a ground-based radar while it is constant for an airborne-Doppler radar, which leads to slightly different algebra. For a ground-based radar, the relationship between the radar-relative azimuth α and the storm-relative azimuth ϕ can be written as

$$\begin{aligned} \sin\alpha &= \frac{\rho}{\delta \cos\beta} \sin\phi + \frac{\omega}{\delta \cos\beta} \sin\gamma, \\ \cos\alpha &= \frac{\rho}{\delta \cos\beta} \cos\phi + \frac{\omega}{\delta \cos\beta} \cos\gamma, \end{aligned} \quad (3)$$

where ω is the distance between the radar and the storm center, δ the distance between the radar and the considered point, and γ the angle between east and the direction of the storm center as "viewed" from the radar (Fig. 1). Hence, using (3), the Doppler velocity in (2) can be written as a modified Fourier analysis of five independent values (D_i) with respect to storm-relative azimuth ϕ and distance δ for each ring at given radius ρ and altitude z :

$$\begin{aligned} V_{\text{DOP}} &= D_0 \left(\frac{\omega}{\delta \cos\beta} \right) + D_{1c} \left(\frac{\omega}{\delta \cos\beta} \cos\phi \right) \\ &\quad + D_{1s} \left(\frac{\omega}{\delta \cos\beta} \sin\phi \right) + D_{2c} \left(\frac{\omega}{\delta \cos\beta} \cos 2\phi \right) \\ &\quad + D_{2s} \left(\frac{\omega}{\delta \cos\beta} \sin 2\phi \right) = \sum_{i=1}^5 [p_i D_i]. \end{aligned} \quad (4)$$

The decomposition in (4) is not mathematically valid when (i) $\delta = 0$, which never happens since measurements are always made at some distance from the radar; (ii) $\beta = 90^\circ$, that is, radar pointing toward zenith; and (iii) $\omega = 0$ (the radar and the storm center are collocated) in which case the tangential velocity is everywhere perpendicular to the radar beam and the T_i components are inaccessible. These limitations did not apply to the considered dataset. Provided that N Doppler velocity measurements at different storm-relative azimuth ϕ are available, the D_i values in the different rings can be determined from the minimization of cost functions J_D (one for each ring):

$$J_D = \sum_{n=1}^N \left\{ \sum_{i=1}^5 [p_i(n)D_i - V_{\text{DOP}}(n)] \right\}^2; \quad \left[\frac{\partial J_D}{\partial D_i} = 0 \right]_{i=1,5}. \quad (5)$$

The relationships between the D_i , $i = 1, 5$ in (4) and the TR_m , $m = 1, 6$ in (2) are

$$\begin{aligned} D_0 &= -\left(\frac{1}{2} \sin \gamma\right) T_{1c} + \left(\frac{1}{2} \cos \gamma\right) T_{1s} + \left(\frac{\rho}{\omega}\right) R_0 \\ &\quad + \left(\frac{1}{2} \cos \gamma\right) R_{1c} + \left(\frac{1}{2} \sin \gamma\right) R_{1s}, \\ D_{1c} &= -(\sin \gamma) T_0 + (\cos \gamma) R_0 + \left(\frac{\rho}{\omega}\right) R_{1c}, \\ D_{1s} &= +(\cos \gamma) T_0 + (\sin \gamma) R_0 + \left(\frac{\rho}{\omega}\right) R_{1s}, \\ D_{2c} &= -\left(\frac{1}{2} \sin \gamma\right) T_{1c} - \left(\frac{1}{2} \cos \gamma\right) T_{1s} + \left(\frac{1}{2} \cos \gamma\right) R_{1c} \\ &\quad - \left(\frac{1}{2} \sin \gamma\right) R_{1s}, \\ D_{2s} &= +\left(\frac{1}{2} \cos \gamma\right) T_{1c} - \left(\frac{1}{2} \sin \gamma\right) T_{1s} + \left(\frac{1}{2} \sin \gamma\right) R_{1c} \\ &\quad + \left(\frac{1}{2} \cos \gamma\right) R_{1s}, \end{aligned} \quad (6)$$

which can be written as

$$\left\{ \sum_{m=1}^6 [q_{im} TR_m] = D_i \right\}_{i=1,5}. \quad (7)$$

It has to be noted that coefficients γ and ω linking the D_i and the TR_m are constant for a given ring. This is not true for the distance ρ to storm center, which varies slightly within each ring. However, ρ can safely be approximated by the value relative to the middle of the considered ring. Since there are less available information (five D_i s) than unknown values (six TR_m s), some assumptions must be made to solve the problem.

To alleviate this difficulty, LJCD choose to neglect the wavenumber-1 components of the radial wind (R_{1c} and R_{1s}). This leads to an overdetermined system of five equations with four unknowns (T_0 , T_{1c} , T_{1s} , R_0), which can be solved in the least squares sense.

We follow here the same approach as in EVTD (Roux and Marks 1996): the availability of L successive radar sequences with different values of γ_1 and ω_1 (resulting from TC motion) allows the six TR_m s to be deduced from a series of L sets of five D_i s, which leads to an overdetermined system of $L \times 5$ equations (6) or (7) with six unknown values. Likewise, for each ring where D_i values are available for L sequences, this is obtained through the minimization of cost function J_{TR} :

$$J_{TR} = \sum_{l=1}^L \left\{ \sum_{i=1}^5 \left\{ \sum_{m=1}^6 [q_{im}(l) TR_m] - D_i(l) \right\}^2 \right\}; \quad \left[\frac{\partial J_{TR}}{\partial TR_m} = 0 \right]_{m=1,6}. \quad (8)$$

For this method to be efficient, two conflicting conditions must however be satisfied.

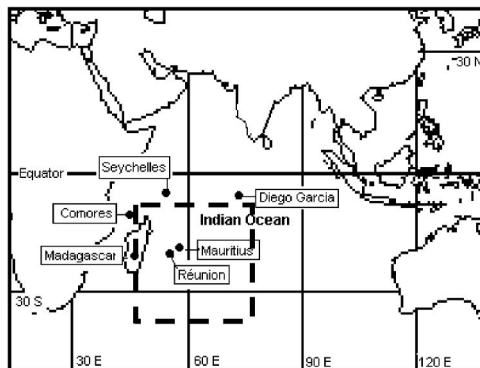
- The values of γ_1 and ω_1 must be as different as possible for the L sets of D_i to be linearly independent, which favors large time intervals (or fast TC motion) between the successive radar sequences.
- This time interval must however be short enough (or storm evolution must be slow) so that the TR_m wind components do not vary too much during the considered period.

Here, series of two or three radar sequences separated by a maximum interval of 1 h have been considered, with associated differences of 5° – 10° in γ and 0–10 km in ω . Tests with simulated data (see the appendix), using the same geometry as the actual radar observation of Tropical Cyclone Dina, have shown that this is sufficient to correctly estimate (rms error $\leq 3 \text{ m s}^{-1}$) the wavenumber-0 and -1 components of the tangential and radial winds. This might not be true in the case of a rapidly evolving storm, but the availability of radar scans at short time intervals (e.g., $\leq 15 \text{ min}$) could alleviate this difficulty.

3. Overview of Tropical Cyclone Dina

Dina formed on 16 January 2002 from a large region of strong convection east of Diego Garcia Island near 8°S , 76°E (Fig. 2; Météo-France 2002). On the 17th, it became a tropical depression, then Tropical Storm Dina while moving rapidly ($>10 \text{ m s}^{-1}$) toward the southwest. On the 18th, the storm was upgraded to tropical cyclone by the Regional Specialized Meteorological Center (RSMC) La Réunion–Tropical Cyclone Centre. Dina displayed a well-defined eye while its propagation speed slowed down to 6 – 7 m s^{-1} . It intensified on the

(a)



(b)

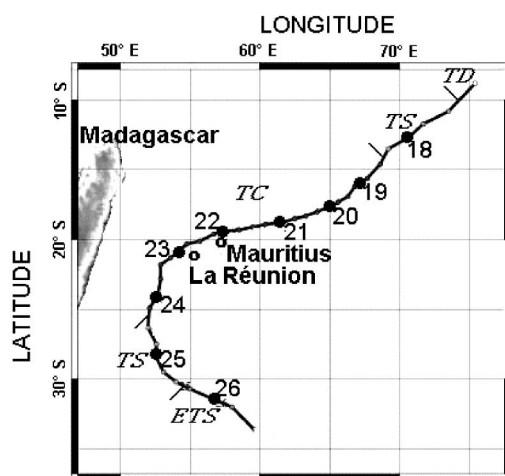


FIG. 2. (a) Map of the Indian Ocean; the dashed-line box denotes the region of interest. (b) Trajectory of Dina in the southwestern Indian Ocean from 17 to 26 Jan 2002; symbols TD, TS, TC, and ETS denote tropical depression, tropical storm, tropical cyclone, and extratropical storm stages, respectively (reproduced with permission from RSMC La Réunion/Météo-France).

19th and during the morning of the 20th with estimated maximum winds of about 70 m s^{-1} and minimum surface pressure of 910 hPa. During the afternoon of 20, 21, and 22 January, surface wind and pressure varied from 60 to 65 m s^{-1} and from 910 to 920 hPa while the direction of motion changed to west-southwestward.

During the night of 21–22 January, Dina passed near Mauritius with a minimum distance of 65 km between the storm center and the northern tip of the island at 2315 UTC on the 21st (Figs. 3a and 3b). At this time, the diameter of the eye was about 85 km and the eyewall was asymmetric with a more intense western side. Dina caused heavy rain (accumulated amount was 350 mm with maxima $>500 \text{ mm}$ over the western part of the island) and strong surface winds (maxima $>60 \text{ m s}^{-1}$).

Dina's eye passed closest to La Réunion ($<130 \text{ km}$) on 22 January between 1000 and 1800 UTC (Figs. 3c and 3d) with a minimum distance of 65 km from the

northern coast at 1330. At this time, the diameter of Dina's eye was about 65 km and the eyewall was relatively symmetric. As deduced from the radar observations discussed below, strong reflectivity values ($>35 \text{ dBZ}$)—and the most intense winds—in the eyewall were found at 40 to 60 km from the storm center, which probably spared La Réunion even more dramatic consequences. Nevertheless, strong winds ($50\text{--}70 \text{ m s}^{-1}$) and heavy rain (rain rates $>50 \text{ mm h}^{-1}$ during more than 12 h, 500–2000 mm accumulated in 72 h), mainly observed over the central high terrain, as well as flooding and high storm surge (6–9 m) in the coastal regions caused major devastation, mostly over the northern half of the island. Dina was one of the strongest cyclones observed in La Réunion in 40 years, and the extent of the damage can also be explained by its relatively slow motion. No casualties were reported, but the estimated cost amounted to several hundred million euros (or U.S. dollars) and it will probably take years before all the damaged infrastructure (roads, water pipes, power and telephone lines, TV and radio transmitters, etc.), forests, farms, industries, buildings, houses, etc., will be rebuilt.

Then on 23 January, Dina's trajectory changed southward and its intensity decreased rapidly: 55 m s^{-1} and 925 hPa on the 23d at 0000 (Figs. 3e,f), 45 m s^{-1} and 955 hPa on the 24th, 25 m s^{-1} and 985 hPa on the 25th, after which it became strongly asymmetric and was carried along the westerly midlatitude circulation as an extratropical depression.

4. Analysis of radar observations

a. Storm propagation

Ten volume scans at 130-km range have been conducted with the Météo-France operational Doppler radar located at 20.89°S , 55.42°E , 743 m MSL (Table 1), every 30 min from 0952 to 1422 UTC 22 January 2002 (Doppler data at 0952 and 1052 were corrupted). The radar antenna and radome were swept away by a strong wind gust at 1450 and no data are available afterward. As seen in Fig. 4, due to the steep topography of La Réunion, with the highest peak—Piton des Neiges—at 3069 m, radar data are not available in the southern quadrant ($115^\circ\text{--}230^\circ$ from the north).

The storm center for each scan was first determined as the geometric center of the eye region, characterized by low ($<20 \text{ dBZ}$) reflectivity values, at the lowest elevation (0.5°). The derived storm track agrees fairly well with the positions of the storm center at 0600 and 1200 derived by Météo-France from a combination of radar and satellite observations. It should be noted that the positions at 1800 and later are from satellite data only, which could explain the relatively short distance between 1422 and 1800. Although the so-determined propagation speed remained nearly constant (6.0 to 6.3 m s^{-1} , in agreement with the large-scale estimate), its direction changed substantially (from 33° to 90° , clock-

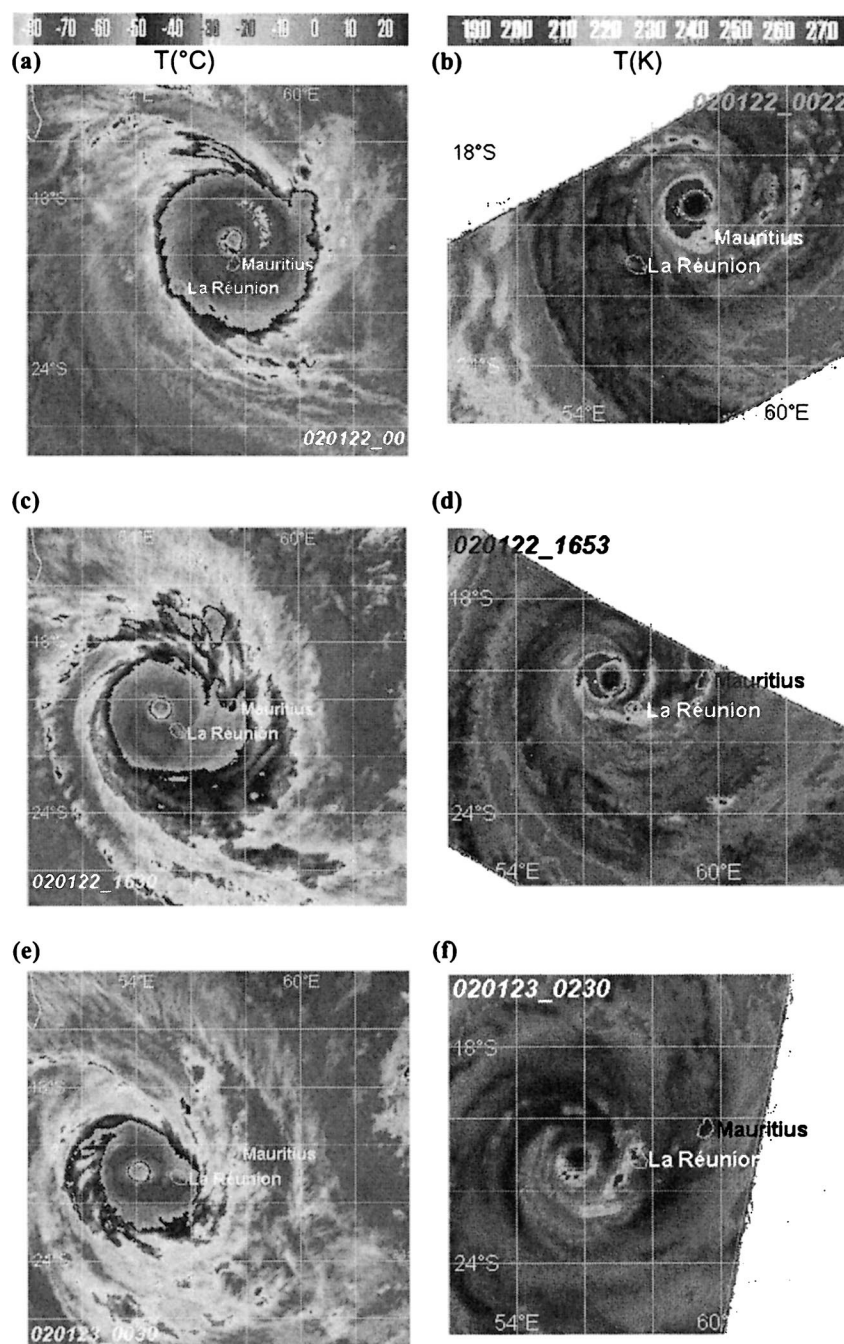


FIG. 3. Passage of Dina near La Réunion: *Meteosat-5* infrared images at (a) 0000 and (c) 1630 UTC 22 Jan, (e) 0030 UTC on 23 Jan; 85-GHz images at (b) 0022 and (d) 1653 UTC 22 Jan (TRMM/TMI), (f) 0230 UTC 23 Jan (DMSP/SSM-I). The gray scales for infrared and microwave images are given (reproduced with permission from Naval Research Laboratory, Monterey Marine Meteorology Division).

wise with respect to north) during the considered period. A similar, although weaker, track deviation was observed from 1800 to 2400 21 January when Dina was less than 150 km from Mauritius where the orography is smoother (highest elevation is at 828 m). Such a situation has already been observed for other tropical cy-

clones passing at a relatively close distance from La Réunion and is probably related to orographic influence. As seen in Fig. 2, Dina turned to a southwestward track later on 22 and 23 January.

Indeed, when a tropical cyclone comes near a mountain range, its track and intensity are influenced by the

TABLE 1. The radar parameters and scanning characteristics of the Météo-France Doppler radar in La Réunion (operated 1992–2002).

Latitude	20.89°S
Longitude	55.42°E
Altitude	743 m (MSL)
Specifications	
Type	Gematronik METEOR 360 AS
Frequency	2.7–2.9 GHz
Peak power	800 kW
Antenna diameter	6 m
Beamwidth (one way, 3 dB)	1.2°
Antenna gain	42 dB
Antenna polarization	Horizontal
Doppler mode	
Pulse repetition frequencies	525 + 700 Hz (dual PRF)
Unambiguous velocity	$\pm 57 \text{ m s}^{-1}$
Pulse length	1 μs
Receiver	Linear
Minimum detectable signal	−106 dBm
Minimum detectable reflectivity	2 dBZ (SNR = 0 dB, range = 50 km)
Clutter suppression	40 dB
Range	130 km (260 range gates of 500-m width)
Azimuthal rotation rate	10° s^{-1}
Azimuthal increment	1°
Elevation angles (°)	0.5, 1.0, 1.5, 2.4, 3.4, 4.3, 5.3, 6.2, 7.5, 8.7, 10.0, 12.0, 14.0, 16.7, 19.5

orography, which makes forecasting a very difficult task. One well-known example is the impact of the Taiwan's Central Mountain Range on approaching typhoons (e.g., Bender et al. 1987; Yeh and Elsberry 1993a,b; Lin et al. 1999, 2002; Wu and Kuo 1999; Wu 2001). As the outer circulation of a TC begins to interact with topography, blocking and deflection of the flow advecting the storm cause zonal deceleration and equatorward deflection, which produce a cyclonic track curvature. This effect increases when the strong cyclonic winds impinge more directly on an orographic barrier. A similar situation probably occurred when Dina, a Southern Hemisphere cyclone following a southwestward track, passed less than 150 km north of La Réunion.

b. Dynamic center and eye rotation

As noted by Lee and Marks (2000), a few kilometers error in the estimated TC center position does not pose serious difficulties in TC track analysis, but it can significantly affect interpretation of the TC wind field in a cylindrical coordinate system. Supposing that the storm center is the geometric center of the eye contour, determined with a radar reflectivity threshold, is therefore not precise enough, and it is necessary to call for a specific algorithm using Doppler velocity data to determine a “dynamic” storm center more closely related

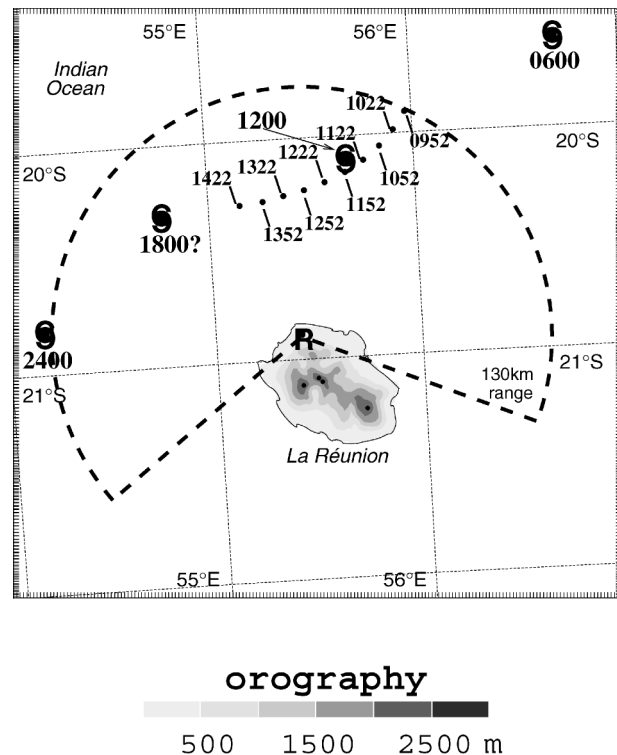


FIG. 4. Location of Dina's center between 0600 and 2400 UTC 22 Jan (the hurricane symbols denote “official” positions issued by Météo-France; dots denote positions derived from radar reflectivity data). The dashed line contour displays the 130-km range of Doppler scans with the Météo-France radar, located in the northern part of La Réunion. Shadings represent increasing altitudes.

to the wind field, for each volume scan. There are, however, many ways of defining such a vortex center: it can be the location of minimum surface pressure, zero wind, maximum vorticity, the center that maximizes the mean tangential wind within an annulus near the radius of maximum wind, or the (potential) vorticity centroid, etc.

Lee and Marks (2000) showed that the simplex method (Nelder and Mead 1965) can efficiently locate the TC center leading to the highest tangential wind within a given annulus for an axisymmetric TC and for asymmetric ones constructed by adding higher wavenumber tangential winds. However, when a wavenumber-1 asymmetric radial wind is present, such a method can converge toward a slightly different location. This is not surprising since, as shown by Willoughby (1992), a displaced TC center yields a spurious wavenumber-1 component in the derived wind field that cannot be distinguished from the actual one.

This difficulty can be deduced from (5), which shows that, from the D_i deduced from single volumetric Doppler data through a GB-VTD analysis [Eq. (4)], it is not possible to estimate the symmetric tangential wind component T_0 , but only a combination of T_0 and the asymmetric radial wind components R_{1c} and R_{1s} :

$$\begin{aligned}
 & -\sin\gamma D_{1c} + \cos\gamma D_{1s} \\
 & = T_0 + \left(\frac{\rho}{\omega}\right) [-\sin\gamma R_{1c} + \cos\gamma R_{1s}] = T_0 R_1. \quad (9)
 \end{aligned}$$

Wind asymmetries in tropical cyclones frequently result from the presence of mesovortices associated with perturbed tangential and radial winds of relatively similar amplitudes. Hence, $T_0 R_1$ cannot always be used as a convergence criterion in the simplex algorithm. Except for vortices with $R_{1c} = 0$ and $R_{1s} = 0$, its amplitude will not always decrease away from the dynamic center where T_0 is maximum, owing to contribution of the actual wavenumber-1 component of the radial wind, and of other components being aliased onto the wavenumber-1 radial wind when the analysis is conducted with a displaced center. An alternative convergence criterion for defining the dynamic storm center can be derived from (5) through

$$\begin{aligned}
 \cos\gamma D_{2c} + \sin\gamma D_{2s} &= \frac{1}{2} [-T_{1s} + R_{1c}] = U, \\
 -\sin\gamma D_{2c} + \cos\gamma D_{2s} &= \frac{1}{2} [T_{1c} + R_{1s}] = V, \quad (10)
 \end{aligned}$$

where U and V are the Cartesian (eastward and northward, respectively) wind components within the considered ring. Hence, we consider the area- and density-weighted mean wind modulus UV , defined as

$$UV = [\langle U \rangle^2 + \langle V \rangle^2]^{1/2} \quad (11)$$

with

$$\begin{aligned}
 \langle U \rangle &= \frac{\sum_{kz=1}^{N_z} \sum_{ir=1}^{N_r} (\Delta_{ir,kz} S_{ir} d_{kz} U_{ir,kz})}{\sum_{kz=1}^{N_z} \sum_{ir=1}^{N_r} (\Delta_{ir,kz} S_{ir} d_{kz})}, \\
 \langle V \rangle &= \frac{\sum_{kz=1}^{N_z} \sum_{ir=1}^{N_r} (\Delta_{ir,kz} S_{ir} d_{kz} V_{ir,kz})}{\sum_{kz=1}^{N_z} \sum_{ir=1}^{N_r} (\Delta_{ir,kz} S_{ir} d_{kz})},
 \end{aligned}$$

where iz and kz are indices for radial distance and altitude, respectively, of the considered ring; $\Delta_{ir,kz}$ is equal to 1 if D_i values can be deduced from GB-VTD analysis in ring ir , kz to 0 otherwise; S_{ir} is the horizontal area of the considered ring; and d_{kz} is the air density at the considered altitude. Here UV can be used as a criterion in the simplex algorithm since, usually, TCs translate approximately with the depth-averaged wind velocity in the inner core region (e.g., Marks et al. 1992; Franklin et al. 1993): UV is a slightly more robust criterion than $T_0 R_1$ since D_{2c} and D_{2s} are uniquely related to U and V in (10), and spurious contributions from other components being aliased onto wavenumber-1 tangential and radial winds will generally lead to larger values of UV . However, it can lead to an incorrect determination of

the dynamic center of a storm moving at some speed, or some angle, with respect to the mean wind. It must also be recognized that higher ($n > 1$) wavenumber tangential and radial wind components can contribute to U and V in (10), as well as in $T_0 R_1$ in (9), and make the determination of the dynamic center more problematic. Likewise, nonuniform data filling, either azimuthally or vertically, would probably have some influence on the determination of $T_0 R_1$ and UV .

Here, to determine the location of the dynamic center of Dina, the simplex algorithm was used with the following characteristics: To form the initial simplex, four points were considered at 3-km distance east, north, west, and south of the “geometric” TC center and a triangle was formed with the points (including the geometric center) leading to the three smallest values of UV . The reflection, expansion, and contraction coefficients were taken as 1.0, 1.0, and 0.5, respectively; 5–15 iterations were needed to find a dynamic TC center leading to $UV < 0.1 \text{ m s}^{-1}$. The obtained results for the eight successive volumic Doppler scans from 1022 to 1422 22 January 2002 are shown in Fig. 5 (the Doppler data at 1052 were corrupted and it was not possible to determine the associated location of the dynamic TC center). The distance between the geometric and dynamic centers varied from less than 2 km (at 1222 and 1352) to about 7 km (at 1022 and 1152). It has to be noted that, although they are not maximum, the area- and density-weighted mean values of $T_0 R_1$ obtained from (9) using the dynamic center are equal to or larger than those obtained when the geometric center is considered, with larger differences for wider distances between the two centers (+4.6 and +1.7 m s^{-1} at 1022 and 1152, respectively).

From the series of reflectivity fields at 1-km altitude in the central part of the storm, it clearly appears that Dina had an elliptical eye with minor and major axes of about 65 and 80 km. Moreover, the eye rotated cyclonically with a period of about 150 min, twice as long as the time necessary for an air parcel to orbit the eyewall (about 75 min at 40-km radius and 55 m s^{-1} speed). This is very similar to the observations of Typhoon Herb in 1996 by Kuo et al. (1999) who found minor and major axes of about 40 and 60 km and a rotation period of approximately 144 min, which they explained as a potential vorticity wave (the generalization of Rossby waves) propagating as a wavenumber-2 asymmetry with a speed slower than the mean flow. Likewise, numerical simulations of Hurricanes Bob in 1991 by Braun (2002) and Bret in 1999 by Nuissier et al. (2004) revealed a wavenumber-2 asymmetry that rotated cyclonically around the storm center at about half the speed of the mean tangential wind, in agreement with the theory for vortex Rossby waves. Unfortunately, due to the relatively poor time resolution (30 min), it is not possible to precisely relate the more complex trajectory of the dynamic center with the rotation of the elliptic eye. Nevertheless, the shift between the geometric and the

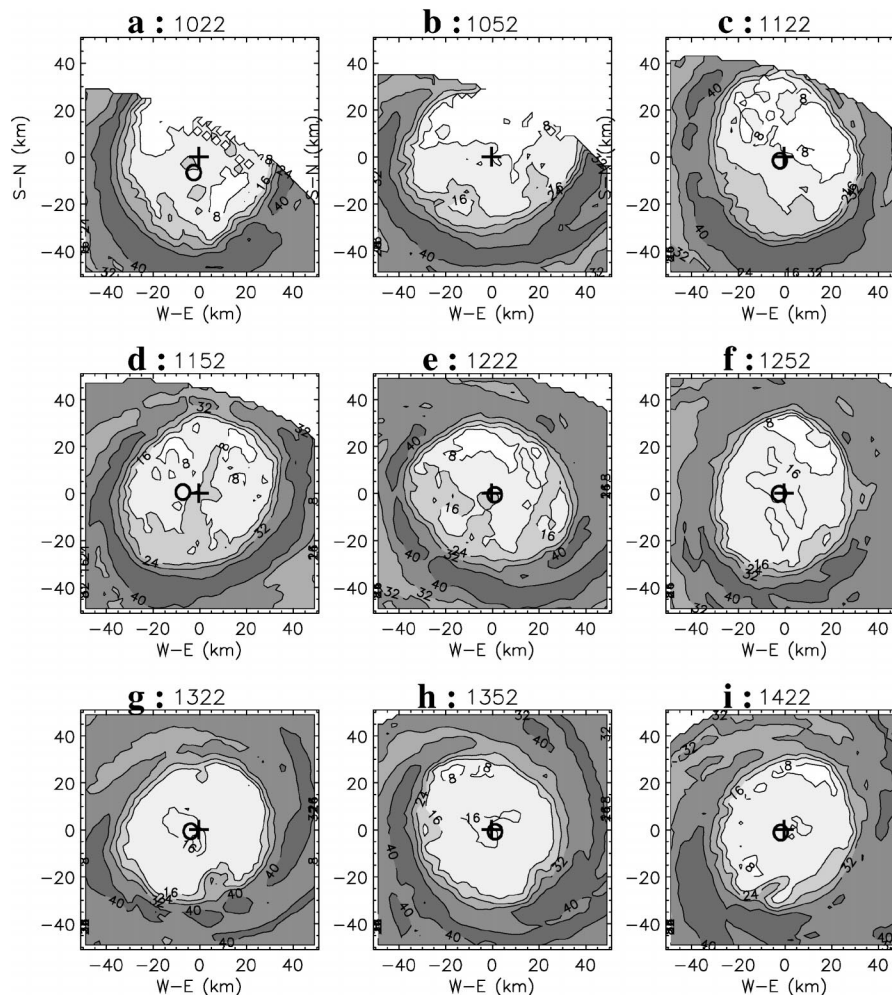


FIG. 5. Horizontal reflectivity fields (step for the contours is 8 dBZ) at 1-km altitude in domains of $100 \text{ km} \times 100 \text{ km}$ centered on the “geometric center” of the storm (denoted by crosses) at (a) 1022, (b) 1052, (c) 1122, (d) 1152, (e) 1222, (f) 1252, (g) 1322, (h) 1352, (i) 1422. Circles denote the position of the “dynamic” center.

dynamic centers seemed to be predominantly along the major axis of the elliptic eye toward the region of highest reflectivity values. Unfortunately, the truncation of the GB-VTD-derived winds at wavenumber 1 prohibits any further analysis of the associated perturbations in the wind field. It must also be outlined that, as discussed in the appendix, the GB-EVTD analysis should alias the probable wavenumber-2 component onto the symmetric and wavenumber-1 fields.

Another interesting feature in Fig. 5 is the presence of precipitation in the lower part of the eye, characterized by reflectivity values greater than 16 dBZ. Although one cannot dismiss the possibility that they could be radar artifacts, these features are similar to those revealed by numerical simulations. In their study of Hurricane Andrew (1992), Liu et al. (1999) found that, below the “eye inversion layer” from 2- and 4-km altitude, a frictionally forced vertical circulation can induce ascent of moist air at the center of the eye. In their

simulation of Hurricane Bret (1999), Nuissier et al. (2004) found that mesovortices developing in the eyewall region were associated with the transport of precipitation inside the eye in the low levels. The reflectivity contours protruding from the internal part of the eyewall toward the eye at 1022, 1322, and 1422 (Figs. 5a,g,i) and the radar echoes near the storm center from 1252 to 1422 (Figs. 5f-i) are very similar to these simulated features.

c. Quality of the GB-EVTD-derived winds

GB-EVTD analyses were conducted with the Doppler data collected in La Réunion on 22 January 2002 for 35 rings of 3-km width around the storm center ($0 < \rho < 105 \text{ km}$) and 30 levels of 500-m depth ($0 < z < 15 \text{ km}$). Three sets of scans were considered: 1022–1122 (intermediate time 1052), 1152–1222–1252 (intermediate time 1222), and 1322–1352–1422 (inter-

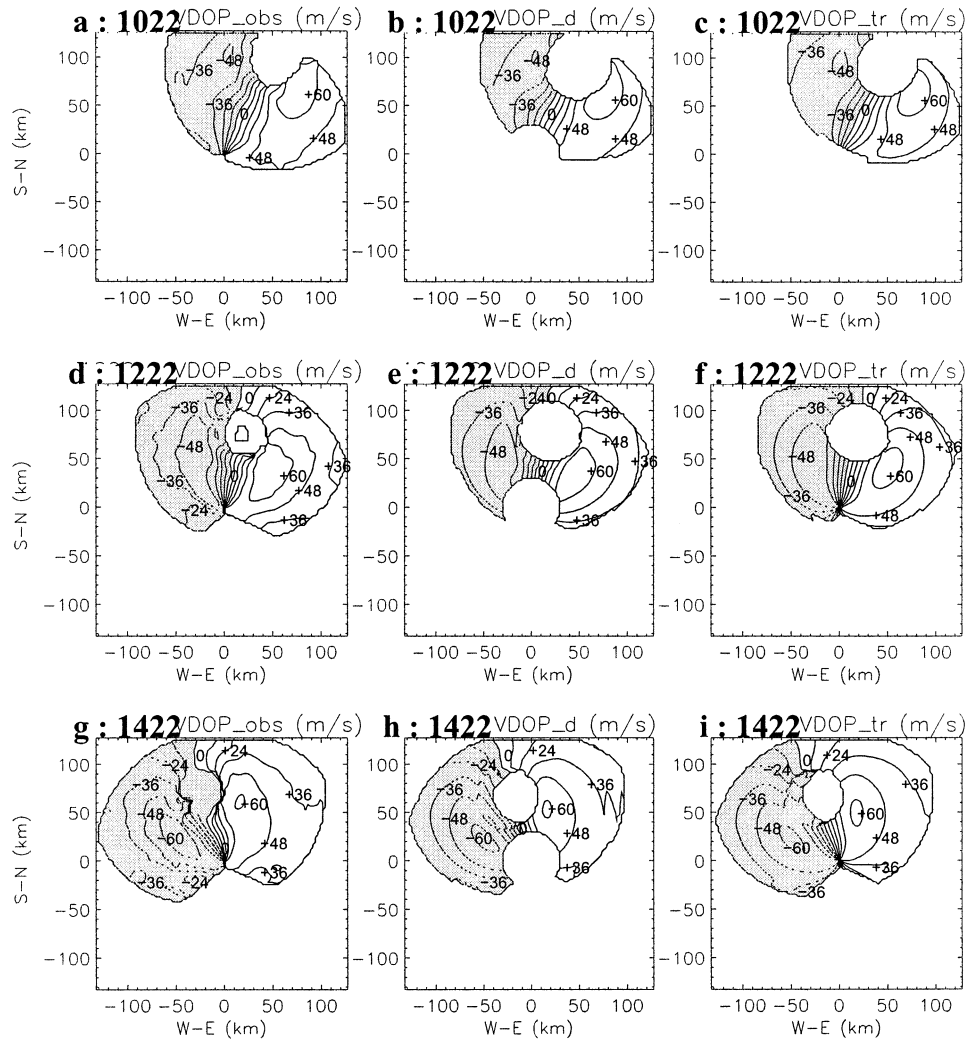


FIG. 6. Observed, GB-VTD-, and GB-EVTD-derived Doppler velocity values at 0.5° elevation, at (a), (b), (c) 1022; (d), (e), (f) 1222; and (g), (h), (i) 1422.

mediate time 1352). It is to be noted that GB-EVTD analysis could be conducted only for those rings where Doppler data were available within an azimuthal sector of more than 180° , which is an important limitation considering the rather large distance (≥ 70 km) between the radar and storm center.

Before discussing the obtained results, it is necessary to verify that the GB-EVTD analysis correctly represents the observed Doppler velocities. Figure 6 shows examples of the observed Doppler velocities, those calculated from (4) with the obtained D_i , $i = 1, 5$ (hereafter referred to as GB-VTD Doppler velocities) and from (2) with the obtained TR_m , $m = 1, 6$ (GB-EVTD Doppler velocities) for 0.5° elevation at 1022, 1222, and 1422. It can be seen that, although the small-scale features are filtered out through the GB-VTD and GB-EVTD analyses, the main characteristics are preserved. The region without data at less than 25 km from the radar for the GB-VTD values is due to the lack of low-

level data throughout the rings, owing to the radar latitude (743 m) and elevation of the first scan (0.5°), which prohibits direct calculation of D_i values at the first level (500 m).

Table 2 gives rms values of the observed, GB-VTD, and GB-EVTD Doppler velocities for each sequence from 1022 to 1422 and the standard deviations of the GB-VTD- and GB-EVTD-derived values with respect to the observed ones. First, it can be seen that the GB-VTD- and GB-EVTD-derived amplitudes are within ± 0.1 and ± 0.8 m s $^{-1}$, respectively, from the observed ones. Second, the standard deviations are 1.9–3.1 and 3.7–6.5 m s $^{-1}$, respectively. Different phenomena can explain the fact that the standard deviations of the GB-EVTD Doppler velocities are slightly greater than the 3 m s $^{-1}$ obtained when dealing with simulated data: contribution of wavenumber-2 and higher tangential and radial wind components, transient features that did not persist during the considered 1-h periods such as local

TABLE 2. Rms values of the observed, GB-VTD- and GB-EVTD-derived Doppler velocity values, and standard deviations of the GB-VTD- and GB-EVTD-derived Doppler velocity values with respect to the observed ones.

Time (UTC)	rms. observed (m s^{-1})	rms.GB- VTD (m s^{-1})	std.GB- VTD (m s^{-1})	rms.GB- EVTD (m s^{-1})	std.GB- EVTD (m s^{-1})
1022	41.8	41.7	1.9	41.6	6.5
1122	42.5	42.4	2.4	41.7	5.9
1152	43.2	43.2	3.1	43.3	4.9
1222	41.9	41.9	2.3	42.0	4.9
1252	41.8	41.7	3.1	41.1	4.6
1322	41.6	41.5	2.6	41.0	3.7
1352	41.1	41.0	2.8	40.7	3.9
1422	41.0	41.0	2.5	41.7	3.7

convective motions and wavelike perturbations propagating around the storm, differences in the orographic influence of the island due to storm propagation, etc.

The larger deviations of the GB-EVTD-derived velocities at 1022 and 1122 result certainly from the fact that only two sequences were considered instead of three for the later times. As discussed in section 2, the smaller standard deviations of the GB-VTD-derived velocities do not imply that more reliable tangential and radial wind components would be deduced from this method. Indeed, GB-VTD provides a more accurate determination of the D_i than GB-EVTD does for the TR_m , but—except in the rare cases where wavenumber-1 components of the radial wind are *actually* equal to zero—only the TR_m have direct physical meaning as wind components, while the D_i s are an incomplete set (<6) of linear combinations.

d. Structure and evolution of the GB-EVTD-derived winds

As deduced from (10), the Cartesian wind components U and V can be derived within each ring from the retrieved D_{2c} and D_{2s} coefficients. Then, the mean values at each level (weighted by the relative surface area of the rings where the analysis is conducted) give a hodograph of the mean wind between 30- (i.e., outside the eye) and 105-km radius from the storm center. For simplicity, Fig. 7 shows the mean hodographs obtained for each group of sequences (intermediate times 1052, 1222, and 1352). As discussed in section 4b, the underlying hypothesis for vortex center determination is that the density-weighted deep-layer mean wind (derived from a GB-VTD analysis of each sequence) should be equal to zero when the Doppler velocities are corrected for propagation speed and are analyzed with respect to the dynamic storm center. Hence, the change in mean wind (crosses in Fig. 7) should reflect the presence of a spatially varying steering flow. As discussed in, for example, Bender et al. (1987), mountainous islands affect the basic flow field, causing changes in track and translation speed, and induce modifications in the

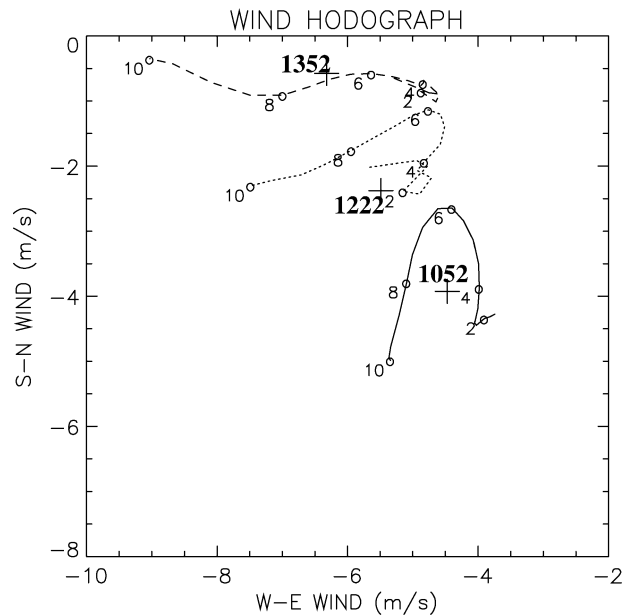


FIG. 7. Hodograph of the mean wind between 30 and 100 km from the storm center and between 2- and 10-km altitude at 1052 (solid line), 1222 (dotted line), and 1352 (dashed line). Crosses indicate storm motion at the different times, deduced from the successive positions of the storm center.

structure of tropical cyclones when they pass nearby. Dina's translation speed remained more or less constant at $6 (\pm 0.2) \text{ m s}^{-1}$, while its track veered from 49° from north at 1052 to 67° at 1222 and 85° at 1352. Such a northwestward deflection of about 4 m s^{-1} in 3 h (or 65 km, see Fig. 4) could have resulted from the force due to a horizontal pressure gradient of about 0.7 Pa km^{-1} from 150° with respect to north, probably caused by the island orography.

The GB-EVTD-derived mean relative flow in Tropical Cyclone Dina veered anticyclonically (counterclockwise in the Southern Hemisphere) with altitude. Between 2- and 6-km altitude, the very weak ($<0.5 \times 10^{-3} \text{ s}^{-1}$) wind shear was southerly at 1052, westerly at 1222, and nearly nonexistent at 1352. Aloft, between 6 and 10 km, it was slightly stronger (0.7 to $0.9 \times 10^{-3} \text{ s}^{-1}$) and northerly at 1052, northeasterly at 1222, and easterly at 1352. This shows that both the large- and the mesoscale wind fields underwent substantial changes when Dina passed near La Réunion. The orientation of the retrieved wind shear differs from the cyclonic ones previously deduced for other storms [e.g., Norbert (1984) by Marks et al. (1992); Hugo (1989) by Roux and Marks (1996); Claudette (1991) by Roux and Viltard (1995)]. However, the hodograph obtained by Reasor et al. (2000) in Hurricane Olivia (1994) also veered anticyclonically with height. It must however be noted that the storm-relative wind intensity and shear in Dina were significantly smaller than those associated with the previous storms.

No GB-EVTD analysis was attempted for reflectivity since (i) it is a logarithmic quantity that cannot be added

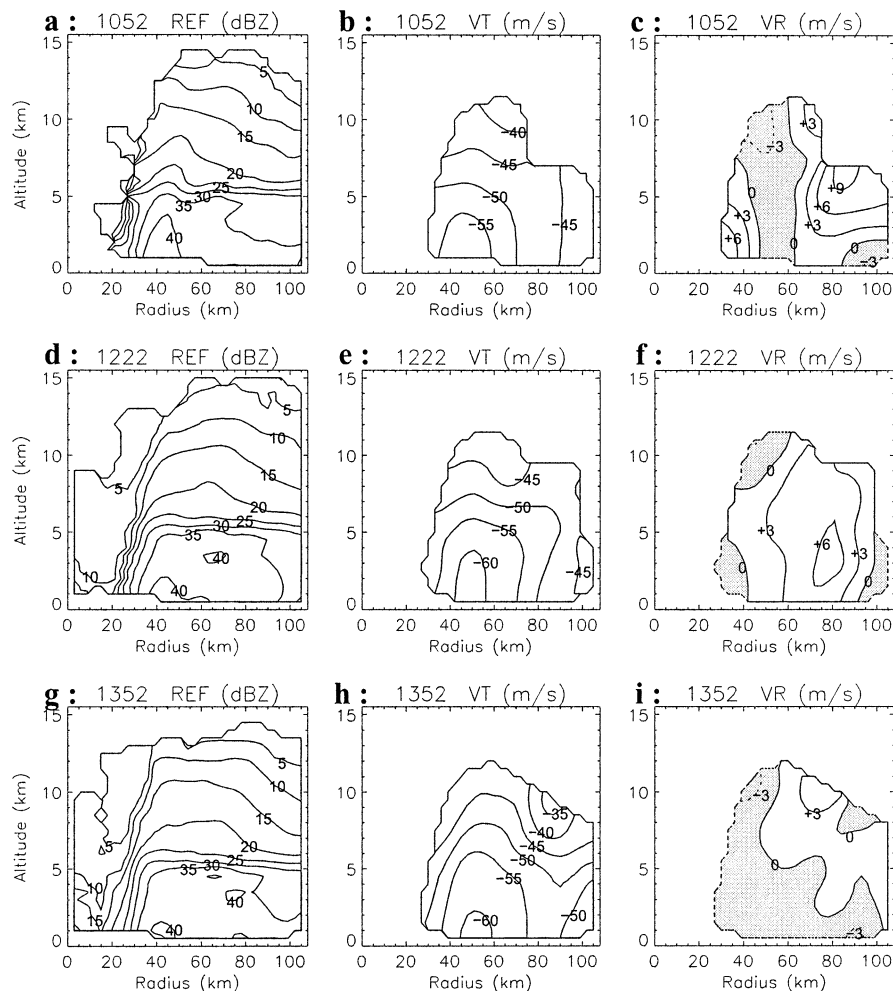


FIG. 8. Mean radial and vertical distribution of (a), (d), (g) radar reflectivity, (b), (e), (h) tangential, and (c), (f), (i) radial wind components for the composite analyses at (a), (b), (c) 1052, (d), (e), (f) 1222, and (g), (h), (i) 1352. Steps for the contours are 5 dBZ for reflectivity, 5 m s^{-1} for tangential wind, and 3 m s^{-1} for radial wind.

and (ii) the negative values, which could result from GB-EVTD analysis of reflectivity-derived precipitation contents (using the same relations as Gamache et al. 1993), would have no physical meaning. Hence, the mean radial and vertical distribution of reflectivity was derived from the azimuthal average of the estimated precipitation content (then transformed in dBZ) for various radii and altitudes around the storm center. Since reflectivity data were not available throughout a 360° storm-relative azimuthal sector for the different radii and altitudes due to range limitation, these mean values differ probably from true azimuthal means, especially at large radii. The observed structure (Figs. 8a,d,g) is characteristic of an intense tropical cyclone with low values (<20 dBZ) in the eye at radii smaller than 30 km, strong (>40 dBZ) and elevated (15 dBZ limit up to 10–11-km altitude) ones in the eyewall region at radii between 30 and 80 km, and slightly weaker and more stratiform values beyond 80 km. With time, the reflec-

tivity maximum (>40 dBZ) near 40-km radius progressively decreased while the reflectivity values beyond 70-km radius intensified slightly. As discussed above, the presence of low, but significant, reflectivity values in the eye could be an indication of a frictionally forced vertical circulation below an eye inversion layer (Liu et al. 1999). Although this hypothesis is not directly supported by the GB-EVTD analysis, as vertical motions were not calculated, it could be related to the intensifying radial mean inflow (Figs. 8c,f,i). The decrease of the 5-dBZ contour from 8 to 6.5 km at 20-km radius from 1222 to 1352, associated with an increase of the 10-dBZ contour from 1–2 to 4–5 km at less than 10 km from the storm center, could be an indication of downward and upward motions, respectively.

The mean circulation is deduced from the GB-EVTD-derived wavenumber-0 (symmetric) tangential T_0 (Figs. 8b,e,h) and radial R_0 (Figs. 8c,f,i) wind components. The strongest tangential winds were collocated with the

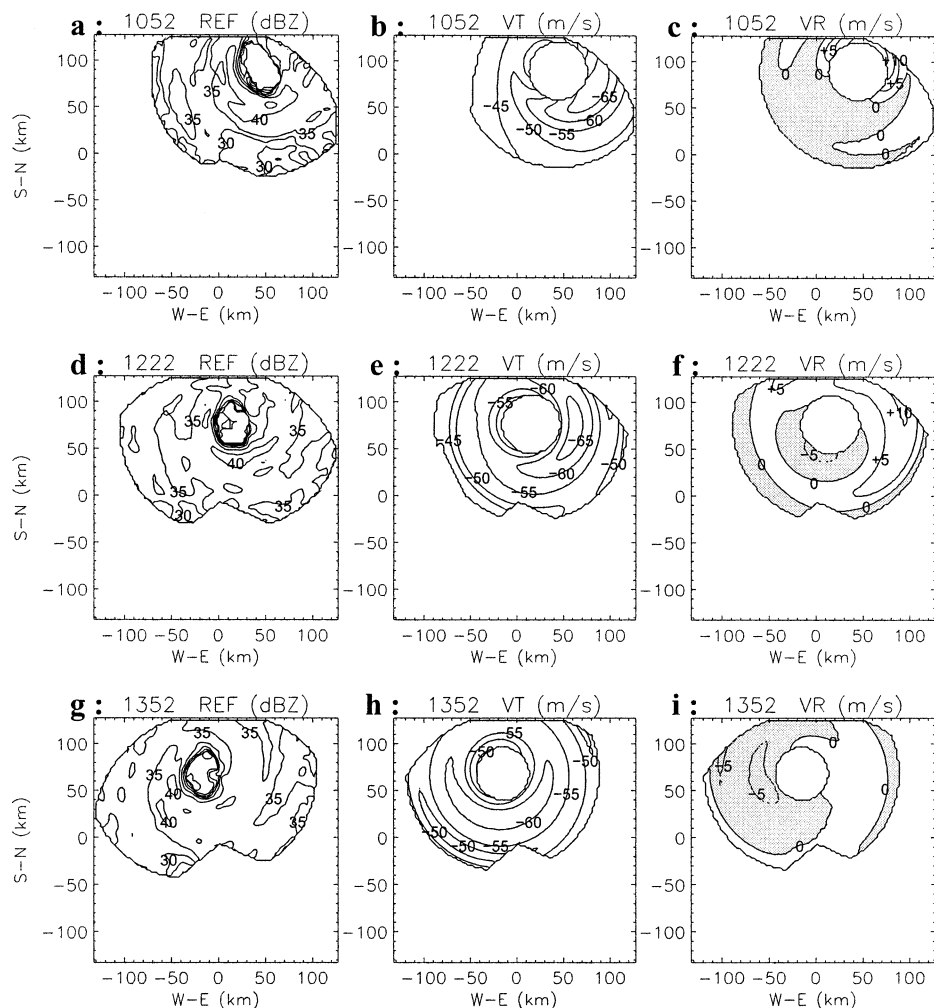


FIG. 9. As in Fig. 8 except for a horizontal cross section at 2-km altitude and 5 m s^{-1} step for radial wind. Coordinate system is positive eastward and northward with its origin at radar location (20.89°S , 55.42°E).

highest mean reflectivity values between 40- and 60-km radii, with maximum values greater than 55 m s^{-1} at 1052 and 60 m s^{-1} at 1222 and 1352, below 3-km altitude. At 500-m altitude, the radius of maximum wind did not change much during the considered period: 48 km at 1052 (59 m s^{-1}), 48 km at 1222 (61 m s^{-1}), and 51 km at 1352 (61 m s^{-1}). Except in the eye region ($\rho < 30 \text{ km}$) where no GB-EVTD-derived winds are available, the mean tangential wind was everywhere greater than 35 m s^{-1} , up to 11-km altitude. Intensity of the tangential wind intensified significantly in the outer part of the domain with the 50 m s^{-1} threshold at 3-km altitude and 70-km radius at 1052, 80 km at 1222, and 100 km at 1352. The mean radial wind was mostly outward (up to $+9 \text{ m s}^{-1}$ at 1052), except at 1352 with weak inflow ($\geq -3 \text{ m s}^{-1}$) in the inner part of the domain. The main inflow feeding the inner core region of tropical cyclones is known to occur in the lowest levels ($< 1\text{--}2 \text{ km}$ altitude; e.g., Tabata et al. 1992), which were

unfortunately not correctly sampled because of radar altitude and first-scan elevation. Inward motions at less than 40-km radius above 8-km altitude could be an indication of downward motions on the inner side of the eyewall.

The horizontal distribution of radar reflectivity and tangential and radial winds at 2-km altitude is shown in Fig. 9. High reflectivity values ($> 40 \text{ dBZ}$) and strong tangential winds ($> 60 \text{ m s}^{-1}$) are found in the eyewall region. Automated surface stations on the northern coast of La Réunion [near (0, 0) in Fig. 9] reported wind maxima $> 50 \text{ m s}^{-1}$ between 1000 and 1600 UTC (gusts $> 60 \text{ m s}^{-1}$ were observed at higher altitude in the mountainous central part of the island), in agreement with the radar-derived values. A notable feature is the changing location and intensity of the maximum wind from 68 m s^{-1} in the east at 1052 to 62 m s^{-1} in the south at 1352. This is correlated with the presence of maximum reflectivity values south of the eye at 1052 and 1222,

then west of it at 1352, in regions of horizontal convergence due to decelerating tangential wind. It can also be seen in Figs. 9b, 9e, and 9h that wind asymmetry decreased significantly with time (at 50-km radius it was 20 m s^{-1} at 1052, 10 m s^{-1} at 1222, and 5 m s^{-1} at 1352). It must be noted that tests with simulated data using the same geometry as the actual radar observations showed that such changes were not related to the propagation of the storm with respect to the radar [i.e., to changing ω and γ in (6); see Fig. 1 and the appendix]. The radial wind shows a mean inflow from the west with some indication of cyclonic (clockwise) turning with time. Such a pattern agrees with the southwestward to westward propagation of Dina (Fig. 2). At 6-km altitude (Fig. 10), relatively similar though weaker features were found: with time, the tangential wind maximum moved from southeast to south-southwest while decreasing slightly and the wind field became more symmetric, radar reflectivity weakened in the southeastern part of the eyewall and intensifies in the southwestern part, and radial inflow moved from southwest to west.

A mean azimuth–height cross section in the eyewall region (radii between 39 and 57 km) is shown in Fig. 11. Again, good correlation appears below 5-km altitude between the reflectivity maxima and the region of horizontal convergence due to inflow and decrease of tangential wind. These regions moved cyclonically (clockwise) by about 90° between 1052 and 1352. Except for the slightly decreasing inflow, not much change is found in the upper levels. A remarkable feature in Dina was the absence of significant azimuthal tilt of the wind and reflectivity contour with height, as was observed in Hurricanes Hugo (1989) (Roux and Marks 1996) and Claudette (1991) (Roux and Viltard 1995).

e. Discussion

The observed evolution of the storm structure probably resulted from its interaction with La Réunion Island. Figure 12 is a very simple sketch displaying the possible influence of the island orography on the cyclonic flow at less than 100 km from the storm center. The island is represented by a 75-km-long (in the NNW–SSE direction) and 55-km-wide (WSW–ENE) ellipse (see Fig. 4), and the airflow is supposed to pass around it. Of course, actual circulation was certainly more complex with the flow passing also over the mountains and into the valleys, and the possible generation of vertically oriented vortices on the lee side of the obstacle with enhanced turbulence and diffusion. In Fig. 12, wind deviation is supposed to result only from two factors: wind intensity and its orientation with respect to the idealized coast.

When Dina was at 100 km or more from the northeastern coast of La Réunion (e.g., composite analysis at 1052), the orographic influence was limited to radial distances ≥ 70 km and storm-relative azimuth between about 170° and 200° . Hence, storm propagation and ki-

nematic structure of the inner core region were probably not very different from the offshore characteristics. As Dina came nearer (e.g., 1222 and 1352), the orographic influence on the cyclonic flow became more important because of the presence of stronger tangential winds in the southeastern quadrant of the eyewall and more perpendicular orientation of the cyclonic flow with respect to the coast, so the region of perturbed wind was wider (radii ≥ 30 km, azimuth between 120° – 140° and 180°).

In these conditions, the cyclonic flow decelerated upstream due to the blocking effect of the elevated terrain in the central part of the island, and it accelerated downstream. Such an effect is observed in Figs. 9b, 9e, and 9h and Figs. 11b, 11e, and 11h with decreasing tangential velocities east of storm-relative azimuth 180° and increasing ones to the west. The extension of the inflow region in the southwestern quadrant (Figs. 9c,f,i; Figs. 11c,f,i) could be an indication that the perturbed flow on the lee side did not return immediately to its original radial distance with respect to storm center. These orographically induced changes of the cyclonic flow, with decreasing northeasterly winds in the southeastern quadrant of the storm and increasing easterly winds in the southern part, could also explain the observed westward deflection of the direction of storm propagation (denoted by arrows in Fig. 12). Although these effects are relatively moderate as compared to the significant mesoscale variations of pressure, wind, and precipitation distribution occurring when a tropical cyclone interact with a larger obstacle, such as Taiwan's Central Mountain Range (e.g., Wu and Kuo 1999), they are of utmost importance to correctly forecast the local impact of a storm approaching a relatively small and densely populated island like La Réunion where evacuation is not possible.

5. Summary and perspectives

The GB-EVTD extension of the GB-VTD technique (Lee et al. 1999; Lee and Marks 2000; Lee et al. 2000) allows us to retrieve the wavenumber-0 and -1 components of tangential and radial wind from a series of ground-based Doppler observations of a tropical cyclone. Minimization of the deep-layer density-weighted mean horizontal wind gives the position of the “dynamic” storm center. Such an analysis of the data collected with the Météo-France operational Doppler radar in La Réunion during the passage of intense Tropical Cyclone Dina at less than 130 km from the island on 22 January 2002 provided useful information on the reflectivity and wind structure and evolution. However, a limitation of the present study was the fact that Doppler measurements were available only twice per hour so that composite GB-EVTD analyses could only be conducted with sets of data collected up to 1 h apart. Reducing the interval between the Doppler sequences would allow shorter du-

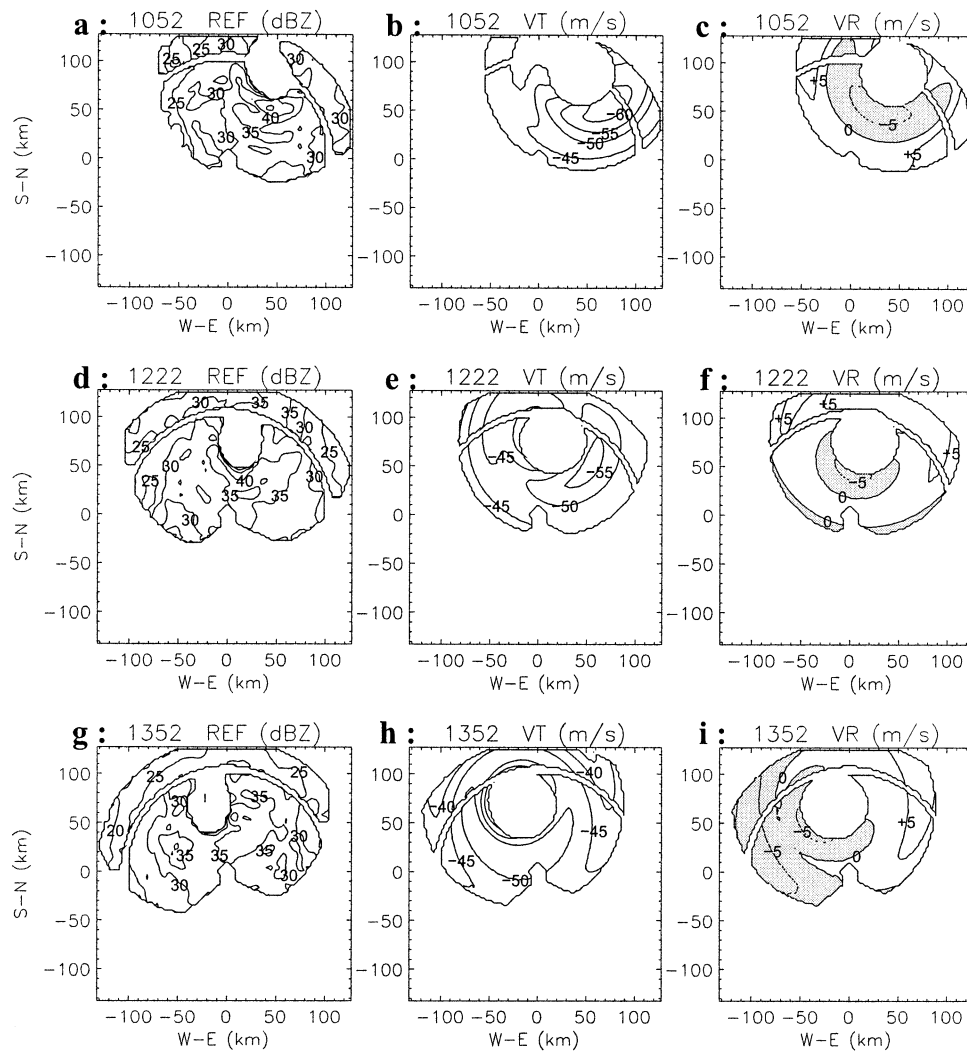


FIG. 10. As in Fig. 9 except for 6-km altitude.

ration composites or more reliable ones based on more numerous sequences [i.e., sets of D_i s in (4)].

From the GB-EVTD-derived winds, it is theoretically possible to calculate the horizontal wind divergence, and through integration of the airmass continuity equation, the vertical velocity (Roux and Marks 1996). However, due to radar altitude, first scan elevation and earth sphericity, the winds retrieved through GB-EVTD in the lowest levels are probably not precise enough for such a task. Availability of scans at less than 0.5° elevation could somewhat alleviate this difficulty, although great care would have to be taken because of possible errors in reflectivity and Doppler velocity arising from part of the radar beam impinging the sea surface. Nevertheless, it would probably be difficult to correctly sample the “frictionally induced” inflow layer due to its relatively shallow depth. Using the thermal wind relation, it would also be possible to estimate the potential temperature

perturbation and surface pressure through the hydrostatic equilibrium (Viltard and Roux 1998).

Although the oversimplified sketch in Fig. 12 gives some clues to explain the observed evolution of Dina’s kinematic structure, more physically consistent studies are needed to more precisely analyze the influence of the mountainous island on storm propagation and intensity. A complementary approach is high-resolution (<2 km horizontally) numerical modeling, which can generate realistic storms (e.g., Liu et al. 1999). Nuissier et al. (2004) have shown that it is possible to initialize a nonhydrostatic nested model with a combination of operational large-scale analysis and a bogus vortex derived from Doppler radar observations. Such a technique will be used to more precisely study propagation, structure, and evolution of Dina on 22 January 2002.

Although it has some limitations, the most important of which being that azimuthal wavenumbers greater than

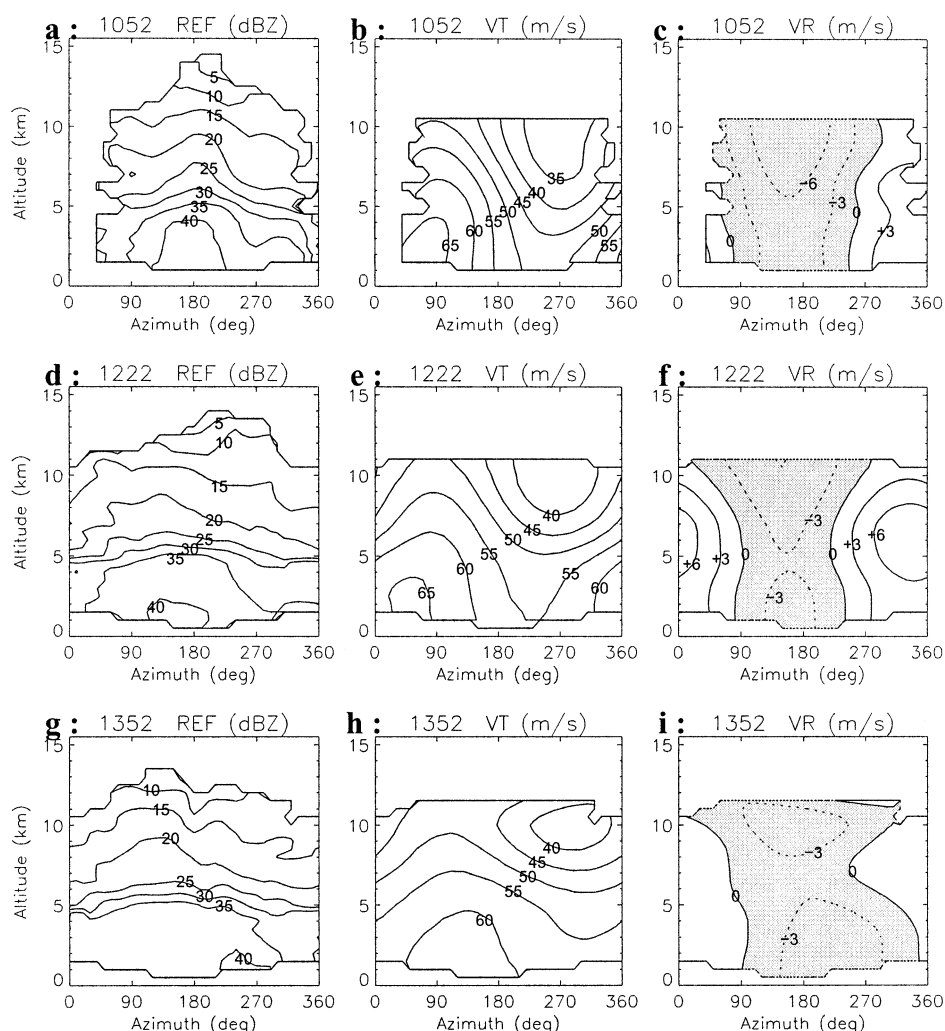


FIG. 11. As in Fig. 8 except for an azimuth–height cross section for radii between 39 and 57 km from the storm center—north: 0°, east: 90°, south: 180°, and west: 270°.

1 cannot be resolved, the GB-EVTD analysis could be of some help for operational purposes when a tropical cyclone approaches an island or a continental area. It is indeed quite difficult to correctly estimate the wind speed based on Doppler velocities only since the projection of the cyclonic wind onto the radar-relative polar coordinate system can lead to major misrepresentation of wind intensity, especially when the strongest winds blow perpendicularly to the radar beam. Through GB-EVTD, it should be possible to more correctly estimate the maximum intensity of the wind and to identify the wavenumber-1 asymmetries in tangential velocity and in radial inflow/outflow. Based on tests with simulated values, a minimum distance of about 10 km between the locations of the storm center during two successive radar sequences is necessary for GB-EVTD to produce useful results. This is equivalent to a minimum time interval of about 2000 s (1000 s) if the storm moves at 5 m s^{-1} (10 m s^{-1}). Of course, data collected at a higher

time rate can be used in the analysis to increase the number of available data. Tests with simulated radar observations from numerical model output should also be conducted to verify whether this condition is compatible with the relatively fast evolution of storms making landfall over a continent or with the structural changes observed when tropical cyclone interact with the steep topography of tropical islands. Nevertheless, even in these cases, GB-EVTD would probably give some useful indication of the average wind intensity. Finally, the scientific and operational interest of GB-EVTD-derived winds would certainly be more important if wavenumber-2 (and, eventually, higher) components could be retrieved. This point certainly deserves further investigation.

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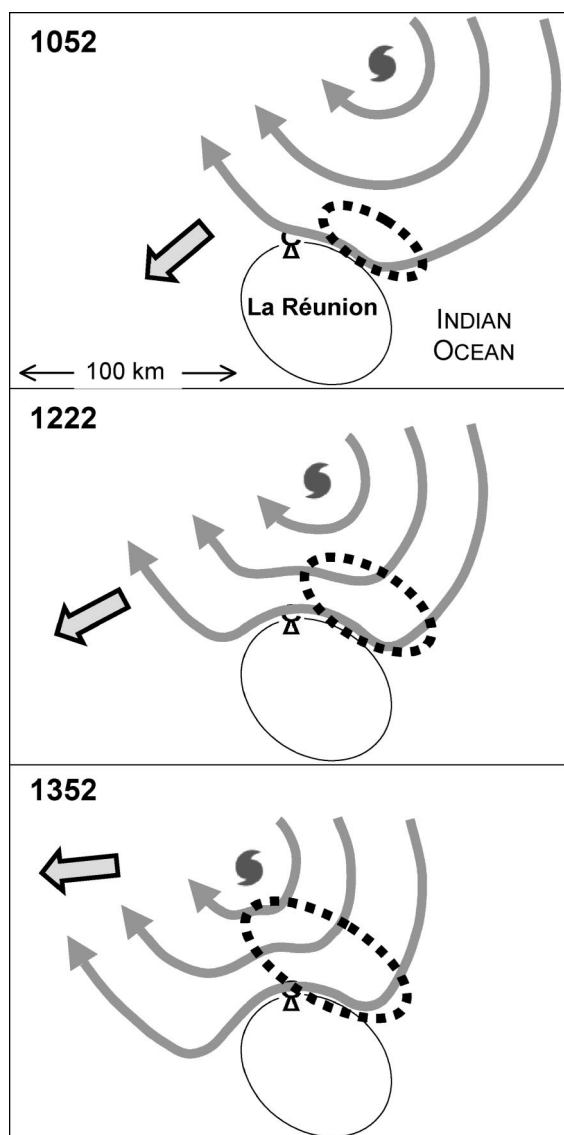


FIG. 12. Sketch of interaction between Tropical Cyclone Dina and La Réunion Island. The light arrows denote the low-level airflow, the dotted-line contour indicates the probable location of the orographically induced convergence, and the contoured arrows represent Dina's motion.

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APPENDIX

Estimated Errors in the GB-EVTD-Retrieved Winds

The GB-EVTD analysis was tested with simulated data using the same characteristics of radar scanning as those relative to the observation of Tropical Cyclone

Dina on 22 January 2002 (see Table 1). The cyclonic wind field was simulated with analytic expressions of the tangential V_T and radial V_R components as

$$V_T(\rho, \phi, z) = F(\rho)G(z) \left[T_0 + \sum_{n=1}^3 [T_{nc} \cos(n\phi) + T_{ns} \sin(n\phi)] \right]$$

$$V_R(\rho, \phi, z) = F(\rho)G(z) \left[R_0 + \sum_{n=1}^3 [R_{nc} \cos(n\phi) + R_{ns} \sin(n\phi)] \right] \quad (\text{A1})$$

with

$$F(\rho) = (\rho/\text{RMW})^{1.5} \exp[1 - (\rho/\text{RMW})^{1.5}]$$

$$G(z) = (1 - z/z_{\text{top}})^{1.5},$$

where ρ is the radial distance from the storm center, ϕ is the storm relative azimuth (0 is eastward, ϕ increases counterclockwise), and z is the altitude above mean sea level. Functions $F(\rho)$ and $G(z)$ are simplified forms of the analytic functions proposed by Holland (1980). The location of the simulated storm was determined by values of the distance ω between the radar and the storm center, and of the angle (or azimuth) γ between the east direction and the line joining the radar and the storm center (see Fig. 1).

A series of GB-EVTD analyses was performed with varying parameters in (A1).

Storm characteristics:

$$T_0 \in [30, 50 \text{ m s}^{-1}],$$

$$(T_{1c}, T_{1s}, R_{1c}, R_{1s}) \in [5, 15 \text{ m s}^{-1}],$$

$$(T_{nc}, T_{ns}, R_{nc}, R_{ns}; \text{ for } n = 2 \text{ and } 3) \in [0, 10 \text{ m s}^{-1}]$$

$$\text{RMW} \in [20, 60 \text{ km}],$$

$$z_{\text{top}} \in [15, 18 \text{ m s}^{-1}].$$

Storm location:

$$\omega \in [20, 110 \text{ km}], \gamma \in [0^\circ, 360^\circ].$$

Storm propagation:

$$U_0 = 5 \text{ m s}^{-1} \cos(\theta);$$

$$V_0 = 5 \text{ m s}^{-1} \sin(\theta);$$

$$\theta \in [0^\circ, 360^\circ].$$

A series of random values within these limits was generated for each parameter, and three simulated Doppler velocity datasets were constructed for three simulated storm locations 30 min (or 9 km with a propagation speed of 5 m s^{-1}) apart. A Gaussian white noise with an rms value of 2 m s^{-1} was added to these Doppler velocity values to simulate radar measurement error. The

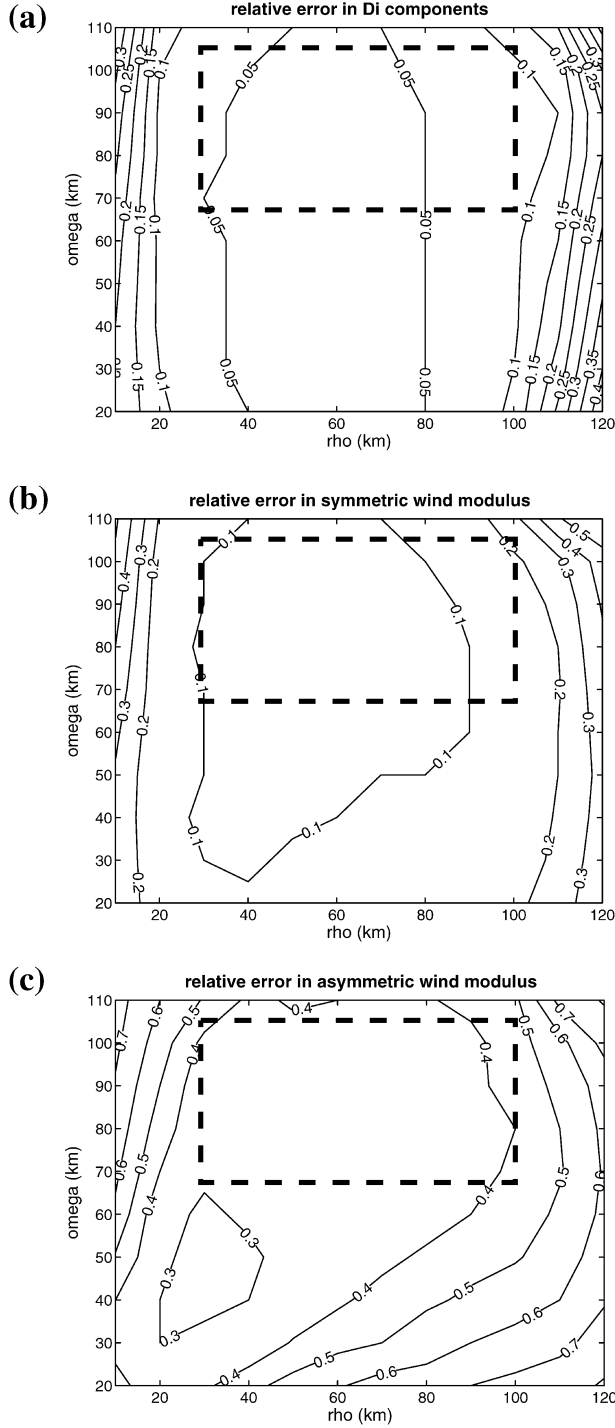


FIG. A1. (a) Relative error (mean quotient between standard deviations and rms values for the series of numerical experiments) of the retrieved D_i , as a function of distance ω between the radar and the storm center, and of radius ρ of the considered ring from the storm center. (b) As in (a) except for the symmetric tangential (T_0) and radial (R_0) wind components. (c) As in (a) except for the asymmetric components (T_{1c} , T_{1s} , R_{1c} , R_{1s}).

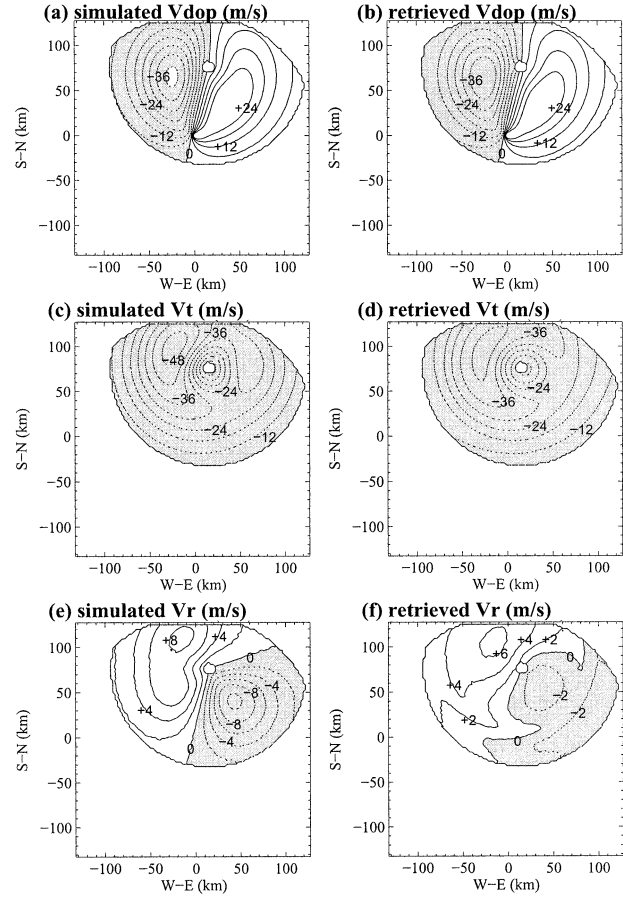


FIG. A2. (a) Simulated Doppler velocity field at the lowest elevation (0.5°), resulting from (c) analytic tangential and (e) radial wind components with contributions from wavenumbers 0, 1, 2, and 3, with $RMW = 45$ km, and $z_{top} = 17$ km; (b) GB-EVTD-derived Doppler velocity, (d) tangential, and (f) radial.

(D_i , $i = 1, 5$) fields were analyzed using GB-VTD within each ring at given altitude and radial distance for each sequence, then the (TR_m , $m = 1, 6$) fields were deduced for each ring from the series of three sets of D_i values.

Figure A1 shows that the relative error (defined as the mean quotient between standard deviations and rms values for the series of numerical experiments) of the retrieved D_i values is small (< 0.1) for all storm locations ($20 < \omega < 110$ km) and for radii ρ between 20 and 100 km from the storm center. No significant change was observed for various radar-relative azimuths γ , which confirms that the GB-EVTD analysis is isotropic. The relative error in the retrieved TR_m fields is slightly larger: it is less than 0.1 for the symmetric tangential T_0 and radial R_0 wind components for $30 < \omega < 10$ km and $30 < \rho < 90$ km, but it is 0.3–0.4 for the asymmetric components (T_{1c} , T_{1s} , R_{1c} , R_{1s}). The largest errors are found at close storm distance ($\omega < 50$ km) and medium to large radii ($\rho > 50$ km), probably resulting from inhomogeneous density and direction of radar measurements in such situations. The domain of

ρ and ω values associated with the observation of Dina on 22 January 2002 (dashed line contour in Fig. A1) was rather favorable for GB-EVTD analyses.

Figure A2a shows the simulated Doppler velocity field at the lowest elevation (0.5°), resulting from analytic tangential (Fig. A2c) and radial (Fig. A2e) wind components with contributions from wavenumbers 0 [$T_0 = 45 \text{ m s}^{-1}$, $R_0 = 2 \text{ m s}^{-1}$], $1[(T_{1c}^2 + T_{1s}^2)^{1/2} = (R_{1c}^2 + R_{1s}^2)^{1/2} = 10 \text{ m s}^{-1}]$, $2[(T_{2c}^2 + T_{2s}^2)^{1/2} = (R_{2c}^2 + R_{2s}^2)^{1/2} = 4 \text{ m s}^{-1}]$, and $3[(T_{3c}^2 + T_{3s}^2)^{1/2} = (R_{3c}^2 + R_{3s}^2)^{1/2} = 2 \text{ m s}^{-1}]$, with RMW = 45 km and $z_{\text{top}} = 17 \text{ km}$. As it can be seen in Fig. A2b, the GB-EVTD analysis was able to correctly reproduce the variations of Doppler velocity in the considered domain. Although the retrieved tangential (Fig. A2d) and radial (Fig. A2f) wind components are truncated to wavenumber 1, the main characteristics of the wind field (location of the maximum and minimum tangential wind, inflow, and outflow regions) are correctly reproduced. There is apparently only limited aliasing of $n > 1$ wavenumbers onto the symmetric and wavenumber-1 fields. However, whereas the intensity of tangential wind is close to the input values, the retrieved radial wind is nearly twice as small as the “actual” one. This limitation must be kept in mind when analyzing the results obtained from real data.

REFERENCES

- Bender, M., R. E. Tuleya, and Y. Kurihara, 1987: A numerical study of the effect of island terrain on tropical cyclones. *Mon. Wea. Rev.*, **115**, 130–155.
- Braun, S. A., 2002: A cloud-resolving simulation of Hurricane Bob (1991): Storm structure and eyewall buoyancy. *Mon. Wea. Rev.*, **130**, 1573–1592.
- Franklin, J. L., S. Lord, S. E. Feuer, and F. D. Marks Jr., 1993: The kinematic structure of Hurricane Gloria (1985) determined from nested analyses of dropwindsonde and Doppler radar data. *Mon. Wea. Rev.*, **121**, 2433–2451.
- Gamache, J. F., R. A. Houze Jr., and F. D. Marks Jr., 1993: Dual-aircraft investigation of the inner core of Hurricane Norbert. Part III: Water budget. *J. Atmos. Sci.*, **50**, 3221–3242.
- Holland, G. J., 1980: An analytic model of the wind and pressure profiles in hurricanes. *Mon. Wea. Rev.*, **108**, 1212–1218.
- Kuo, H.-C., R. T. Williams, and J.-H. Chen, 1999: A possible mechanism for the eye rotation of Typhoon Herb. *Mon. Wea. Rev.*, **127**, 1659–1673.
- Lee, W.-C., and F. D. Marks Jr., 2000: Tropical cyclone kinematic structure retrieved from single Doppler radar observations. Part II: The GBVTD-simplex center finding algorithm. *Mon. Wea. Rev.*, **128**, 1925–1936.
- , F. D. Marks Jr., and R. E. Carbone, 1994: A technique to extract real-time tropical cyclone circulation using a single airborne Doppler radar. *J. Atmos. Oceanic Technol.*, **11**, 337–356.
- , B. J.-D. Jou, P. L. Chang, and S.-M. Deng, 1999: Tropical cyclone kinematic structure retrieved from single Doppler radar observations. Part I: Interpretation of Doppler velocity patterns and the GBVTD technique. *Mon. Wea. Rev.*, **127**, 2419–2439.
- , B. J.-D. Jou, P. L. Chang, and F. D. Marks Jr., 2000: Tropical cyclone kinematic structure retrieved from single Doppler radar observations. Part III: Evolution and structure of Typhoon Alex (1987). *Mon. Wea. Rev.*, **128**, 3982–4001.
- Lin, Y.-L., J. Han, and D. W. Hamilton, 1999: Orographic influence on a drifting cyclone. *J. Atmos. Sci.*, **56**, 534–562.
- , D. B. Ensley, S. Chiao, and C.-Y. Huang, 2002: Orographic influences on rainfall and track deflection associated with the passage of a tropical cyclone. *Mon. Wea. Rev.*, **130**, 2929–2950.
- Liu, Y., D.-L. Zhang, and M. K. Yau, 1999: A multiscale numerical study of Hurricane Andrew (1992). Part II: Kinematics and inner core structure. *Mon. Wea. Rev.*, **127**, 2597–2616.
- Marks, F. D., Jr., R. A. Houze Jr., and J. F. Gamache, 1992: Dual-aircraft investigation of Hurricane Norbert. Part I: Kinematic structure. *J. Atmos. Sci.*, **49**, 919–942.
- Météo-France, 2002: Dina—Cyclone tropical intense à la Réunion (21–24 Janvier 2002). Météo-France, Direction Interrégionale de la Réunion, 64 pp. [Available from Météo-France, Direction Interrégionale de la Réunion, Le Chaudron BP 4, 97491 Sainte-Clotilde Cedex, France.]
- Nelder, J. A., and R. Mead, 1965: A simplex method for function minimization. *Comput. J.*, **7**, 308–313.
- Nuissier, O., R. F. Rogers, and F. Roux, 2004: Numerical simulations of a mature tropical cyclone: Hurricane Bret on 21–22 August 1999. *Quart. J. Roy. Meteor. Soc.*, in press.
- Reasor, P. D., M. T. Montgomery, F. D. Marks Jr., and J. F. Gamache, 2000: Low-wavenumber structure and evolution of the hurricane inner core observed by airborne dual-Doppler radar. *Mon. Wea. Rev.*, **128**, 1653–1680.
- Roux, F., and N. Viltard, 1995: Structure and evolution of Hurricane Claudette on 7 September 1991 from airborne Doppler radar data. Part I: Kinematics. *Mon. Wea. Rev.*, **123**, 2611–2639.
- , and F. D. Marks Jr., 1996: Extended velocity track display (EVTD): An improved processing method for Doppler radar observations of tropical cyclones. *J. Atmos. Oceanic Technol.*, **13**, 875–899.
- Tabata, A., H. Sakakibara, M. Ishihara, K. Matsuura, and Z. Yanagisawa, 1992: A general view of Typhoon 8514 observed by dual-Doppler radar—From outer rainbands to eyewall clouds. *J. Meteor. Soc. Japan*, **70**, 897–917.
- Viltard, N., and F. Roux, 1998: Structure and evolution of Hurricane Claudette on 7 September 1991 from airborne Doppler radar data. Part II: Thermodynamics. *Mon. Wea. Rev.*, **126**, 281–302.
- Willoughby, H. E., 1992: Linear motion of a shallow-water barotropic vortex as an initial value problem. *J. Atmos. Sci.*, **49**, 2015–2031.
- Wu, C.-C., 2001: Numerical simulation of Typhoon Gladys (1994) and its interaction with Taiwan terrain using the GFDL hurricane model. *Mon. Wea. Rev.*, **129**, 1533–1549.
- , and Y.-H. Kuo, 1999: Typhoons affecting Taiwan: Current understanding and future challenges. *Bull. Amer. Meteor. Soc.*, **80**, 67–80.
- Yeh, T.-C., and R. L. Elsberry, 1993a: Interactions of typhoons with the Taiwan orography. Part I: Upstream track deflection. *Mon. Wea. Rev.*, **121**, 3193–3212.
- , and —, 1993b: Interactions of typhoons with the Taiwan orography. Part II: Continuous and discontinuous tracks across the island. *Mon. Wea. Rev.*, **121**, 3213–3233.