Mesoscale dynamics of a deepening secondary cyclone in FASTEX IOP16: Three-dimensional structure retrieved from dropsonde data

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SUMMARY

The mesoscale dynamics of a secondary cyclone sampled during the field phase of the Fronts and Atlantic Storm-Track EXperiment (FASTEX) Intensive Observation Period (IOP) 16 are documented using airborne dropsonde data. The method proposed to recover the mesoscale analytical three-dimensional (3-D) dynamic fields from the dropsonde measurements is evaluated using simulated fields, airborne in situ measurements, and airborne Doppler radar measurements. These simulations and comparisons indicate that the method is able to resolve the mesoscale structures within the studied secondary cyclone both in the regions of clear air and precipitation. It appears, however, that the most accurate description of the 3-D dynamic fields in both these regions would be obtained by combining Doppler and dropsonde measurements.

The studied secondary cyclone is a fast moving frontal wave that developed on the trailing cold front of a parent low situated over Greenland. Using the dropsonde-derived 3-D fields, the mesoscale flows involved in the organization of the secondary cyclone are described. The potential-vorticity and equivalent-potential-vorticity fields (deduced from the basic analytical fields) indicate that the atmosphere is roughly neutral to moist slantwise instability. This configuration is known to significantly promote frontogenesis. The vertical configuration of the jets and circulations within this secondary cyclone are compared with the coupling–uncoupling scheme of Shapiro. A slight coupling is found between the thermally indirect circulation and the ageostrophic circulation. The obtained coupling is, however, not as developed as in the ideal case described by Shapiro, which may explain the small upward velocities in our case.

Airborne Doppler radar and dropsonde measurements are combined so as to infer the multi-scale processes involved in the mesoscale and convective-scale organization of the Atlantic secondary cyclones and to evaluate the new emerging theoretical interpretations of secondary cyclogenesis.

KEYWORDS: Aircraft observations Atmospheric dynamics Cyclogenesis Frontal wave

1. INTRODUCTION

The observational study of secondary cyclogenesis has recently become a major issue, in response to the recent emergence of new theoretical approaches to diagnose this phenomenon and associated problems, especially on the 1000 km scale (see the recent review by Parker 1998). It has been shown primarily that the semi-ageostrophic theory of frontogenesis (Sawyer 1956; Eliassen 1962; Hoskins and Bretherton 1972) was providing a simple but realistic description of atmospheric fronts. In addition, recent efforts were made to provide an instability theory of frontal cyclogenesis at the 1000 km scale (Schär and Davies 1990; Joly and Thorpe 1990), as provided long ago by Charney (1947) and Eady (1949) for the larger-scale cyclogenesis. These new theoretical approaches (e.g. Bishop and Thorpe 1994a,b; Joly 1995; Thorncroft and Hoskins 1990) have led to new theoretical interpretations of cyclogenesis. It has been postulated that cyclogenesis at low levels could involve a variety of mechanisms (see Joly et al. 1997 for further details), but that its subsequent development involves only one mechanism, namely a baroclinic interaction with upper levels. Nevertheless, although these new hypotheses seem to be supported by a few recent studies (e.g. Rivals et al. 1998), it has become a priority to gather and analyse appropriate observational datasets in order to evaluate the realism of these new hypotheses. In this context, an international field project, the Fronts and Atlantic Storm-Track EXperiment (FASTEX, Joly et al. 1997), was designed. The field phase of the FASTEX took place during January and February

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1997 in the North Atlantic Ocean and provided the required observations (Joly et al. 1999).

Detailed scientific objectives of this experiment are summarized in Joly et al. (1997). With regard to cloud systems of cyclones, the major objectives of FASTEX are to test the previous theoretical hypotheses against observations and to advance our understanding of the multi-scale interactions involved in the organization and evolution of mature or developing secondary cyclones. At mesoscale and convective-scale, the cyclone organization is indeed known to be affected by various kinds of activity that still need to be explained, such as the existence of rainbands (e.g. Lemaître and Scialom 1992; Browning et al. 1997) and the break-up of frontal zones into line segments or mesoscale vortices (Browning and Roberts 1996; Neiman et al. 1993). The processes associated with these mesoscale characteristics involve complex interactions between diabatic and dynamical processes which may occur within cloudy air or at the rear part of the cyclone (e.g. Browning et al. 1995).

Along with the evaluation of the current theoretical interpretations of secondary cyclogenesis, the determination of the multi-scale processes involved in the mesoscale and convective-scale organization of the secondary cyclones is the second major scientific objective of our investigations related to FASTEX. To achieve these objectives, we intend to take advantage of the unprecedented observational dataset collected during FASTEX so as to obtain a description as accurate as possible of the three-dimensional (3-D) kinematic and thermodynamic fields at mesoscale and convective-scale.

Documentation of the structure and physics of precipitation in the mature or deepening stage of FASTEX cyclones is achieved in a 500 × 800 km² domain called the Multiscale Sampling Area (hereafter referred to as the MSA), as explained in the FASTEX Operations Plans (Jorgensen et al. 1996). This area is mainly instrumented by research aircraft (NOAA* P-3, NCAR† Electra and UK C-130), which were focused on the in situ precipitation structure, cloud microphysics and airflow circulation associated with secondary cyclogenesis. Furthermore, within the MSA, the co-ordinated aircraft flight plans were designed such that the collected data could be analysed in different ways. At full resolution (thanks to the Doppler radars), convective-scale structures such as rainbands can be described, while at a coarser resolution the mesoscale structure of the whole cyclone can be obtained. The two Doppler radar equipped aircraft (NOAA P-3 and NCAR Electra) are devoted to the documentation of the structure of the 3-D wind and precipitation fields within these mesoscale precipitation regions. However, as explained by Joly et al. (1997), a complete dynamical understanding of the secondary cyclones must include observations in the precipitation-free regions that are not sampled by the radars. For this purpose, the UK C-130 and the dropsonde capability have been deployed in the MSA so as to map for the first time the clear-air dynamic structure of the secondary cyclones in three dimensions.

As a result, the most accurate 3-D kinematic and thermodynamic fields in both the regions of clear-air and precipitation would likely result from a combination of Doppler and dropsonde measurements, as proposed recently by Montmerle and Lemaître (1997).

The other major interest from processing these dropsonde data is that 3-D fields of thermodynamic parameters can be recovered from direct measurements which is not the case when Doppler radar data are used. This approach is original in the sense that 3-D quantities can be retrieved for the first time from dropsonde measurements thanks to the unusually high resolution of dropsonde sampling performed over such a large

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area (the MSA) during FASTEX. However, before merging the Doppler and dropsonde measurements and interpreting the whole dataset quantitatively, the agreement between the data sources and their physical consistency needs to be assessed. This evaluation is carried out in the present paper. A side objective related to the obtaining of these validated 3-D mesoscale fields would be to validate mesoscale numerical models.

In this paper, a new analysis technique for retrieval of 3-D fields from the dropsonde measurements is presented (section 2) and evaluated using simulated fields having the characteristics of a secondary cyclone (section 3). Comparisons are conducted with airborne in situ and Doppler observations in common areas (section 4). Application of the method to the dropsonde dataset collected during the Intensive Observing Period IOP16 of FASTEX (17 February 1997) is presented in section 4, where the dropsonde-derived fields are analysed in order to document the mesoscale structure of the secondary cyclone sampled during IOP16 as revealed by the dropsonde measurements.

2. PRINCIPLE OF THE 3-D RETRIEVAL

(a) Analytical formulation of the retrieval method

The mathematical formulation of this analysis consists of writing all the variables to be retrieved (that is, the three wind components, pressure, temperature, and relative humidity) as a product of three expansions in terms of series of orthonormal functions, each of these expansions depending in turn upon each spatial coordinate. Thus this method relies upon exactly the same principle as the Multiple Analytical DOPpler (MANDOP) analysis (Scialom and Lemaître 1990, hereafter referred to as SL), devoted to the retrieval of the 3-D wind field from Doppler measurements. As in the MANDOP analysis, the orthonormal functions (Legendre polynomials in the present case) are expanded up to a given order.

The coefficients of the corresponding analytical expansion are then obtained for each variable \( V_i \) by minimizing in the least squares sense the difference between the analytic form of the variable and the corresponding dropsonde measurement \( U_{ij} \) (that is, the two horizontal wind components, pressure, temperature, and relative humidity) for all the experimental points (denoted \( e \)) obtained from all \( j \) dropsondes \( (j = 1, 2, \ldots) \) in the retrieval domain. The corresponding condition for each variable may be written in the following form:

\[
P_{\text{obs}}(i) = \sum_j \sum_e (V_i - U_{ij})^2 \text{ minimum.} \tag{1}
\]

Concerning the thermodynamic fields, we note that no additional physical constraints are introduced in the minimization procedure, since the thermodynamic quantities are measured directly by the dropsondes. On the other hand, since dropsondes do not measure the vertical wind component, \( w \), this component cannot be recovered unless physical constraints are added to the previous minimization procedure. Therefore, the anelastic approximation of the airmass continuity equation is introduced in the minimization procedure, as well as a ground-level boundary condition for \( w \), using the same formulation as in SL. In the context of FASTEX, orography is assumed to be negligible. Hence, the vertical wind component is supposed to be nil at ground level. It is to be noted that this variational formulation with physical constraints corresponds to the weak-constraints case proposed by Sasaki (1970). This estimation of constrained least-squares parameters is also clearly described in geophysical data analysis textbooks.
such as Menke (1989). The general condition to be minimized may finally be written as:

\[ P' = \sum_i P_{\text{obs}}(i) + \lambda_1 \sum_d \left( \nabla \cdot \mathbf{V} - \frac{w}{H} \right)^2 e^{-\frac{2(i-x)}{\tau}} + \lambda_2 \sum_g (w)^2 \min \]  

where \( d \) is the total number of grid elements in the whole retrieval domain, \( g \) the number of grid elements at ground level, \( \mathbf{V}(u, v, w) \) the 3-D wind vector, \( H \) the scale height for density variations, \( \lambda_1 \) and \( \lambda_2 \) are weighting factors that depend on the strength of the constraint, and \( \tau \) a reference altitude (taken as the ground level). Evaluation of the weighting factors is carried out following the same standard deviation approach as SL.

(b) Correction for advection

The basic hypothesis of the proposed analysis method is that quasi-instantaneous 3-D fields can be obtained for each quantity measured by the dropsondes, despite the fact that the dropsondes are not launched simultaneously. The consequence of such a bold hypothesis is the need to determine a global displacement speed of the phenomenon that minimizes the internal evolution of the storm during the sampling time (this ‘moving frame of reference’ concept was initially discussed and applied by Gal-Chen 1982). Once this advection speed is known, the sampled cyclone is assumed to be stationary in this storm-relative moving frame. Within this moving frame of reference, we then seek an analytic form of the instantaneous 3-D wind field at time \( t_0 \) (for example the mid-time of the sampling sequence). To find the coefficients of the analytic form, one must compare it with the dropsonde measurements. However, these measurements are not all collected instantaneously at time \( t_0 \). In order to correct for this time shift between each measurement, the comparison with the analytic form is done by displacing the analytic form by \( (-C_x(t-t_0), -C_y(t-t_0)) \), where \( (C_x, C_y) \) represents an advection speed vector in conventional Cartesian coordinates. Therefore correction for advection simply consists in minimizing the difference between a measurement at a time \( t \) and the corresponding analytic field taken at the location it occupies at the same time \( t \).

The main requirement to enable correction for advection is thus to estimate an appropriate global advection speed. Let us indeed recall that a whole set of dropsonde data is collected within a rather large time interval (about 5–6 hours) using the FASTEX dropsonde sampling strategy at mesoscale. The stationarity hypothesis in the moving frame of reference may thus not be satisfied with this experimental design. This type of problem has been investigated in an oceanographic context by Matthews (1997), who used a bilinear interpolation method to recover mesoscale ocean features from a quasi-synoptic sampling of a developing ocean front. In this particular study and with this type of analysis method, significant errors were found to arise from the internal evolution of the mesoscale features. Hence, this possible source of contamination will be studied in the simulations of section 3, where we estimate the impact on the retrieval of advection speed errors and of the potential existence of significant internal evolution.

(c) General properties of the method

As already discussed in SL, Dou et al. (1996), Protat et al. (1998) and Montmerle and Lemaitre (1997), the analytical approach used in this analysis benefits from several advantages, that can be summarized as follows.

(i) It avoids the numerical errors that would be associated with a spatial discretization (in particular of the anelastic approximation of the airmass continuity equation).

(ii) Any source of data or physical equation complementary to the Doppler or dropsonde measurements can be easily introduced in the minimization procedure.
(iii) The analytic formulation is characterized by natural filtering and interpolating properties. As for MANDOP, the cut-off wavelength \( \lambda_{c2} \) of the method in a given direction \( x \) results from the grid resolution \( \Delta x \), the domain size \( D_x \), and the order of expansion \( n \) of the analytic functions. As explained in SL, no structure with a wavelength smaller than \( \lambda_{c1} = 2\Delta x \) can be resolved according to Shannon's theorem. On the other hand, a wave along each axis defined by a Legendre polynomial expanded up to order \( n \) reaches zero at most \( (n - 1) \) times along the domain size, which implies a minimal wavelength \( \lambda_{c2} \) of \( 2D_x/(n - 1) \). Parameters \( n, D_x, \) and \( \Delta x \) are thus chosen in the three directions of space such that \( \lambda_{c1} < \lambda_{c2} \). Let us recall that the associated minimum observable scale is \( l_c = 0.26\lambda_{c2} \) (SL).

(iv) All other fields computed from the retrieved analytic fields that would help interpretation of the cyclone dynamics would in turn be expressed in an analytic form (e.g. divergence, vorticity, potential vorticity and equivalent potential vorticity etc.). In particular, an analytic representation of the ageostrophic wind components (crucial to evaluate the theoretical interpretations of cyclogenesis using the FASTEX dataset) can be derived from these retrieved analytic fields using the AVAG methodology (Lemaître et al. 1994).

3. APPLICATION TO SIMULATED DATA

This section describes numerical tests carried out to assess the robustness of the retrieval of the 3-D dynamical and thermodynamical fields from dropsonde data. First, a test field must be constructed. For this purpose, a mathematical expression of the horizontal wind components \( u \) and \( v \) is defined. This mathematical expression is a mesoscale cyclonic circulation, whose parameters are chosen in such a way that it has the overall characteristics of the explosive secondary cyclones observed during FASTEX in terms of horizontal and vertical wind magnitudes, horizontal and vertical scales of vorticity, horizontal divergence, and pressure gradients. The maximum values for \( u, v \) and \( w \) are approximately 40, 40 and 1 m s\(^{-1}\), respectively, while the simulated horizontal divergence and vertical vorticity peak at \( 2 \times 10^{-4} \) s\(^{-1}\) and \( 3 \times 10^{-4} \) s\(^{-1}\), respectively. The horizontal structure of the 3-D wind and pressure fields are shown at 1.5 km altitude in Fig. 1(a). A vertical cross-section through the associated vorticity maximum (Fig. 1(b)) shows that typical features of secondary cyclones, such as the classic low-level and upper-level vorticity anomalies, are well represented using this mathematical formulation. This wind field is assumed in sections 3(a) to 3(c) to be stationary in a moving frame of reference characterized by an advection speed vector \((C_x, C_y)\).

The vertical wind component of this test field is then derived mathematically by introducing \( u \) and \( v \) in the anelastic approximation of the air-mass continuity equation, and using the boundary condition \( w = 0 \) at the ground. Finally, the pressure field of this test field is derived by introducing the mathematical expression of the three wind components in the anelastic approximation of the two horizontal momentum equations using a stationarity assumption \( (\partial u/\partial t = \partial v/\partial t = \partial w/\partial t = 0) \) in the reference frame moving at \((C_x, C_y)\). Note that these horizontal momentum equations are only used to generate a pressure field physically consistent with the simulated 3-D wind field. To avoid confusion, it must be clearly stated here that such a physical consistency would not be necessary in the present case, since in our method the thermodynamic fields are retrieved in real cases from direct dropsonde thermodynamic measurements, without any assumption related to the physical consistency between the wind, pressure, and temperature solutions.
Each dropsonde observation samples a much broader bandwidth of spatial scales than can be resolved by the present method. A crude way to add this kind of 'input noise' to the previous perfect mathematical solution is to treat this input noise as a white-noise process characterized by a given error variance for each of the measurements. Convective-scale retrievals within secondary cyclones generally indicate that the peak updraught and downdraught magnitudes are about 1 m s\(^{-1}\).

It may thus be considered that the impact of scales smaller than the scale resolved by the analysis results in a maximum contribution of 1 m s\(^{-1}\) to the vertical wind component. Then, rough calculations using the anelastic approximation of the continuity equation applied to a convective cell 10 km wide and 5 km deep also lead to a maximum horizontal wind variation of about 1 m s\(^{-1}\). This value of 1 m s\(^{-1}\) has thus been chosen as the error amplitude for the white-noise process added to the mathematical expression of the three wind components. Corresponding error amplitudes for the pressure field are not easy to obtain. Horizontal pressure fluctuations due to convection have been imposed as 0.5 hPa. Finally, the test fields used for all the simulations of section 3 are the addition of the previous mathematical expression of 3-D wind and pressure and of a white-noise process characterized by its error amplitude.

Errors are estimated throughout this section using the classical root-mean-square (r.m.s.) difference, \(rmsd\), between the mathematical expression \(A_{sim_i}\) (without the white noise) of a given field \(A\) and the retrieved field \(A_{ret_i}\) averaged over the whole domain:

\[
\text{rmsd}(A) = \left\{ \frac{1}{M} \sum_{i=1}^{M} (A_{ret_i} - A_{sim_i})^2 \right\}^{1/2}
\]

(3)

where \(M\) is the number of valid points in the retrieval domain. The accuracy of the first-order derivatives of the retrieved fields will also be presented in the following simulations. For the sake of simplicity only the r.m.s. errors of the horizontal divergence, \(div\), and the vertical component of vorticity, \(\zeta\), will be presented. Their mathematical expressions (the 'true' mesoscale fields) have been derived by hand from the expression of the wind components.

In what follows, different sources of contamination that are thought to be the most likely to occur are investigated, namely the impact of an irregular data coverage
(section 3(a)), of measurement errors (section 3(b)), of an error in the advection speed of the moving frame of reference (section 3(c)), and of neglecting the internal evolution of the storm in this moving frame of reference (section 3(d)). In all the simulations, the size of the mesoscale domain is $700 \times 700 \times 7$ km$^3$, with a grid spacing of $10 \times 10 \times 0.3$ km$^3$. This corresponds to the mesoscale domain that has been generally sampled using the dropsonde sampling strategy defined for FASTEX.

The order of expansion of the Legendre polynomials is 5 in the horizontal directions and 7 in the vertical, corresponding to minimum observable scales of 90 km and 600 m, respectively (see section 2(c)).

(a) Impact of irregular data coverage on the retrieval

The effect of a naturally irregular data coverage is evaluated in this section. The dropsondes are considered to be launched simultaneously along a simulated aircraft trajectory, with a realistic sampling strategy (the 'lawnmower' pattern used during FASTEX, Jorgensen et al. (1996)). Only 44 irregularly spaced dropsondes are considered along this trajectory. Using this framework, r.m.s. differences between the simulated and retrieved pressure, $u$, $v$, $w$, $du$, and $\zeta$ fields are 0.26 mb, 0.07 m s$^{-1}$, 0.07 m s$^{-1}$, 0.013 m s$^{-1}$, $1.4 \times 10^{-6}$ s$^{-1}$ and $1.4 \times 10^{-6}$ s$^{-1}$, respectively. Note that these errors arise from both the white-noise process added to the perfect fields, and the irregular and sparse data spacing. This result shows that such a dropsonde sampling is sufficient to recover the simulated mesoscale features in the absence of other sources of contamination, and especially the mesoscale features related to the first-order derivatives of the wind. For mesoscale applications, the r.m.s. error obtained for the pressure field would seem rather large. However, inspection of the horizontal distribution of errors for pressure (not shown) reveals that absolute errors vary horizontally from 0 up to 0.3 mb on a representative horizontal scale of roughly 600 km, which corresponds to a small error in the horizontal pressure gradient. This maximum error in pressure would correspond (using a semi-geostrophic approximation) to an error in the horizontal ageostrophic wind component of 0.4 m s$^{-1}$, which is reasonable.

(b) Impact of additional measurement errors on the retrieval

In this section, the impact of dropsonde measurement errors is evaluated. Real noise of dropsonde measurements has been estimated for the LORAN-C technology by Thorpe and Clough (1991) as approximately 1 m s$^{-1}$ and 2 mb for the wind and pressure measurements, respectively. These values will thus be used in the simulations as the standard deviations of our simulated measurement errors, although during FASTEX many of the dropsondes used Global Positioning System technology instead of LORAN-C.

A Gaussian noise, which is generally thought of as a relatively good approximation of a measurement error, is added to the basic simulated field described previously (the mathematical expression plus the white noise), with a standard deviation ranging from 0 to 3 m s$^{-1}$ for the wind field and from 0 to 3 mb for the pressure field. Corresponding r.m.s. differences are given in Figs. 2(a) and (b). From these figures, it is seen that there is only a slight increase of the r.m.s. differences for all the parameters (and especially for the first-order derivatives), which shows that the retrieval of the 3-D wind and pressure fields is not degraded much when noise is added to the simulated values. This expected behaviour results from the natural filtering of the analytical formulation discussed in SL.
(c) **Impact of an advection error on the retrieval**

Figure 3 shows the C-130 and P3 aircraft trajectories (dashed and solid lines, respectively) flown within the IOP16 cyclone, superimposed on METEOSAT images in the infra-red channel. The ground-relative P3 aircraft trajectory is shown in Fig. 3(a). The advection speed of the IOP16 cyclone (estimated in real-time by the numerical models during the field phase of FASTEX) was approximately 28 m s\(^{-1}\), 35° north of east. Using this advection speed, the aircraft trajectories are ‘unfolded’ in the moving frame of reference, leading to the flight pattern of Fig. 3(b).

As explained in section 2(b), the determination of an advection speed of the storm that minimizes its internal evolution is crucial to the assumption of stationarity of the
Figure 3. METEOSAT image in the infra-red channel of the FASTEX IOP16 secondary cyclone at 0900 UTC 17 February 1997. Also displayed in this figure is the mesoscale retrieval domain used in sections 3 and 4 (large square) and the associated ($x$, $y$) coordinate system, the direction of advection of the cyclone (bold arrow), and the NOAA P3 and UK C-130 aircraft trajectories (continuous and dashed lines, respectively) (a) in the ground-relative frame, and (b) in the storm-relative moving frame, using an advection speed given in the text.
retrieved fields in the moving frame of reference. This section tests the impact of an error in both the magnitude and direction of the global advection speed vector. The same sampling as in section 3(a) is used (44 dropsondes), and a time lag of 6 hours between the first and last dropsonde measurement is introduced. The correct position of the dropsondes in the retrieval domain is then determined in the storm-relative moving frame using the 'true' advection speed of the storm. The 3-D simulated fields are still taken to be stationary in the storm-relative frame.

The impact on the retrieved fields of an error in the advection speed is given in Fig. 4(a). From this it is seen that the errors in all the parameters are increasing almost exponentially, up to 1.2 mb, 0.8 m s\(^{-1}\), 1.3 m s\(^{-1}\), 0.025 m s\(^{-1}\), 5.5 \times 10^{-6} \text{ s}^{-1} and 1.3 \times 10^{-5} \text{ s}^{-1}, for the pressure, \( u \), \( v \), \( w \), \( \text{div} \), and \( \zeta \) fields, respectively, for an error of magnitude 6 m s\(^{-1}\) in the advection speed, which is an extreme case. The r.m.s. errors associated with a directional error in the advection speed vector (Fig. 4(b)) are also significantly increasing, leading to error values up to 1.2 mb, 5 m s\(^{-1}\), 2.5 m s\(^{-1}\), 0.08 m s\(^{-1}\), 2.4 \times 10^{-5} \text{ s}^{-1} and 2 \times 10^{-5} \text{ s}^{-1}, for the pressure, \( u \), \( v \), \( w \), \( \text{div} \), and \( \zeta \) fields, respectively, and for a 20° directional error.

The maximum r.m.s. differences obtained for the kinematic fields still appear quite reasonable, since they represent a 10 to 15% relative error in the simulated wind components and first-order derivatives. Concerning the 1.2 mb r.m.s. error in pressure found in response to a 6 m s\(^{-1}\) error in the advection speed or a 20° directional error, inspection of the horizontal distribution of this error (not shown) shows maximum errors of up to 2 mb for a representative horizontal scale of 600 km. This maximum error in pressure would correspond (using the semi-geostrophic approximation) to an error in the horizontal ageostrophic wind component of 2.8 m s\(^{-1}\), which, although not very accurate, still appears reasonable.

\[(d) \quad \text{Impact of an internal evolution of the 3-D wind field on the retrieval}\]

As discussed previously, a significant internal evolution of the 3-D fields during the sampling of the mesoscale domain by dropsondes could be a major source of errors, since the retrieved fields could not be reasonably considered as instantaneous fields. The effect of internal evolution is thus estimated using the settings of section 3(e), but with time-dependent simulated fields in the storm-relative moving frame using a given growth rate of the three wind components during the sampling time. This growth rate is applied to each component \(UU = (u, v, \text{or } w)\) as follows:

\[
UU'(t) = UU \cdot [1 + \beta(t - t_0)]
\]

where \( t \) is the current time, and \( t_0 \) a reference time (mid-time of the sampling). Using this procedure, the wind components are not changing uniformly in the horizontal and vertical (as is the case in real situations), since the growth rate is a function of the velocity magnitude itself. This implies that both the magnitude and shape of the fields are changing. The growth rate is controlled by the value of the \( \beta \) parameter, leading to a (100\( \beta \))% variation per hour of the wind components.

Hence, for a 6-hour sampling, a growth rate of 10% per hour would generate a 60% modification of the wind components, which clearly represents an extreme case. Indeed, rough calculation using the gradient-wind balance approximation indicates that this 10% per hour growth rate would correspond to a deepening rate of 18 mb in 6 hours, which is much larger than the typical criterion for explosive cyclogenesis (24 mb in 24 hours, Sanders and Gyakum 1980).
Figure 4. (a) Root-mean-square differences between retrieved and simulated horizontal wind components (m s\(^{-1}\); left-hand axis): \(u\) (black squares), \(v\) (black circles), pressure (diamonds; mb; left-hand axis), vertical wind component \(w\) (triangles; m s\(^{-1}\); right-hand axis), horizontal divergence and vertical component of vorticity (white squares/circles; \(1 \times 10^{-4}\) s\(^{-1}\); right-hand axis) as a function of an error in the magnitude of the storm advection velocity vector. (b) as (a) but as a function of an error in the direction of the advection velocity vector.

Figure 5 shows the r.m.s. differences between the retrieved wind components and the same components simulated at the beginning of the sampling. A quasi-linear increase of the r.m.s. differences as a function of the growth rate is observed for the three components, leading to maximum values of 5.5, 2.8, and 0.11 m s\(^{-1}\) for \(u\), \(v\) and \(w\), respectively, and of \(2.2 \times 10^{-5}\) s\(^{-1}\) and \(3.2 \times 10^{-5}\) s\(^{-1}\) for the \textit{div} and \(\zeta\) fields, respectively, in the extreme configuration for which a 60% modification of the wind components occurs during the sampling time. Considering the IOP16 secondary cyclone studied in section 4, the deepening phase was characterized by a 18 mb reduction of
the surface pressure over 24 hours, which, using the gradient-wind balance hypothesis, corresponds to a maximum growth rate of 2.5%, leading to errors of about 1.3 m s$^{-1}$ in the horizontal wind components and of less than $10^{-5}$ s$^{-1}$ for the $div$ and $\zeta$ fields. Thus, this implies that the obtained mesoscale features may be trusted provided that there is a reasonable internal evolution of the storm occurring in the moving frame of reference. In other words, the stationarity hypothesis is a reasonable assumption except for extreme ‘bomb-like’ deepenings.

4. Mesoscale dynamics of the IOP16 secondary cyclone

The purpose of this section is to document the mesoscale dynamics of the secondary cyclone sampled during FASTEX IOP16. Indeed, only few observational studies of these Atlantic cyclones at mesoscale are available in the literature. A qualitative interpretation is carried out using the 3-D kinematic and thermodynamic fields retrieved by the analysis method, described in section 2, from the dropsonde measurements collected within this secondary cyclone. From these dropsonde-derived fields, the 3-D fields of diagnostic quantities such as the potential vorticity, $PV$, and equivalent potential vorticity, $PV_e$, are derived analytically to help this preliminary interpretation of the mesoscale dynamics of the IOP16 cyclone. However, before interpreting those fields, there is a great opportunity to perform quantitative comparisons between the dropsonde-derived 3-D wind field and other independent data sources in the MSA (in situ measurements collected onboard the P3 aircraft, and airborne Doppler radar measurements). These comparisons are conducted in section 4(a).

(a) Intercomparison between the dropsonde-derived fields and other data sources

First, r.m.s. differences between observations and retrieved quantities have been computed so as to evaluate how well the analysis approximated the original observations. These r.m.s. values are 3 m s$^{-1}$ and 6° for the magnitude and direction of the horizontal wind; 0.4 K, 0.6 K and 0.8 K for temperature, potential temperature and
equivalent potential temperature, respectively; 0.7 mb for pressure and 9% for relative humidity. These differences result from the fact that the retrieved fields are compared with measurements which include the scales resolved by the analysis and also the scales and measurement errors filtered out by the analysis.

At the flight-level of the P3 aircraft, the dropsonde-derived 3-D wind field and in situ measurements collected onboard the P3 aircraft can be compared. The main value of such a comparison is that very different spatial scales are resolved in the dropsonde-derived fields and the individual in situ measurements. Indeed, the dropsonde-derived fields are characterized by a minimum observable horizontal scale of 90 km, while the in situ wind measurements involve a large bandwidth of spatial scales. However, this comparison does not involve different time-scales for the two data sources, since they are almost coincident in time.

Figure 6 shows the absolute in situ horizontal wind vectors superimposed at the 1.5 km level on those retrieved from the dropsonde measurements. For display purposes, the in situ measurements shown are 10 km averaged values along the trajectory. A good overall agreement is obtained between the two independent estimates, with a progressive deceleration of the horizontal airflow from the lower to the upper part of the domain. Some smaller-scale structures and changes in direction of the in situ airflow are, however, not captured by the dropsonde-derived field. This is naturally due to the fact that the dropsonde-derived 3-D airflow is much more filtered than the in situ measurements. The respective magnitudes of the in situ and dropsonde-derived horizontal wind components are compared more quantitatively in Fig. 7 for a given straight-line flight track (the first long leg from the right in Fig. 6). The in situ measurements have previously been filtered in order to smooth the small-scale fluctuations of the signal. A global increase in magnitude of the $u$ and $v$ components along the leg is seen in both estimates of
the horizontal airflow (Fig. 7). Maximum differences between the two estimates are approximately 1.5 m s$^{-1}$, which is relatively small given the different resolutions and derivation methods used to access this horizontal airflow.

Much more can be deduced by comparing the dropsonde-derived 3-D wind field with the 3-D wind field retrieved from Doppler radar observations, using the same order of expansion of the polynomials in both cases (5 in the horizontal directions and 7 in the vertical, corresponding to minimum observable scales of 90 km and 600 m, respectively). It must be noted that using the same order of expansion does not imply at all that the Doppler-derived and dropsonde-derived wind components should be similar. Although both instruments sample roughly the same bandwidth of spatial scales, the resolution of the radar data throughout the retrieval domain (roughly 150 m along each radial direction) is far better than the resolution corresponding to the 33 dropsondes (spaced approximately 70 km apart on average) used to recover the 3-D wind field within the same retrieval domain. The analytic form used in MANDOP to recover the 3-D wind field is thus much more constrained by the radar measurements than is the analytic form used in the dropsonde case.

The second important value of this intercomparison is related to the advection problem, which has been identified as the main source of errors in the simulations of section 3. In both cases (dropsonde and Doppler radar), the aircraft trajectories must be positioned using the advection speed of the optimal moving frame of reference (see section 3(c)). If this optimal moving frame of reference is not well-estimated, then the error in the positioning of the dropsonde and radar measurements is similar. However, the vertical wind component is estimated in the radar case (using MANDOP) from measurements very close in space and time, while in our method this vertical wind
Figure 8. Frames (a) and (b) horizontal cross-sections of the dropsonde- and Doppler-derived 3-D airflow, respectively, at 1.5 km altitude; (c) and (d), similar but for 3 km altitude. All frames show wind vectors in the storm-relative moving frame, with reference vectors to the right of (b) and (d). Frames (b) and (d) also show radar reflectivity in dBZ (see key). Also shown in (a) are the locations of the vertical cross-sections of Figs. 9 and 12 (AB), and Fig. 11 (CD).

component is derived from few measurements, generally wide apart in space and time. Indeed, in the dropsonde case the local horizontal divergence (from which the vertical motions are retrieved) is estimated from dropsonde measurements obtained during two successive legs, that is to say with a time shift ranging from 30 minutes to two hours; while in the radar case this local divergence is estimated from measurements which are generally collected during the same leg and thus with a much smaller time shift. Thus, an error in the advection speed leads to wrong positioning of the dropsondes, and thus to an error in the estimate of horizontal divergence. This implies that the retrieval of the vertical wind component will be much more sensitive to an error in the advection speed in the dropsonde case than in the radar case.

The horizontal wind vectors retrieved from the dropsonde and Doppler measurements are given in Figs. 8(a) and (c), and 8(b) and (d), respectively. These fields are given in the storm-relative moving frame, using the advection speed of the phenomenon given in section 3(c). At 1.5 km altitude (Figs. 8(a) and 8(b)), both fields exhibit a well-defined cyclonic circulation in the western part of the domain. The south-eastern
branch of the circulation (see arrow pointing north in Fig. 8) is characterized by a southwesterly flow that gets progressively diffluent at the exit of the circulation, with easternmost and westernmost parts of the flow that turn anticyclonically and cyclonically, respectively. Similar magnitudes are found for the dropsonde- and Doppler-derived horizontal wind vectors. The north-western part of the vortex (Fig. 8(a) and 8(b)) is characterized by a north-easterly flow that progressively wraps cyclonically around it. Again, magnitudes and changes in direction of the dropsonde- and Doppler-derived flows are comparable. A slight shift in position of the circulation centre is, however, found between the two fields of Figs. 8(a) and 8(b). This can be due to two factors, mainly: (i) the better resolution of the Doppler measurements that are consequently able to provide a more detailed description of the convective-scale dynamics than the dropsonde measurements; and/or (ii) an error in the advection speed estimate, to which the 3-D retrieval using dropsondes would be more sensitive than MANDOP, as discussed previously.

Inspection of the dropsonde- and Doppler-derived horizontal wind vectors at 3 km altitude (Figs. 8(c) and 8(d), respectively) again shows good agreement between the two estimates, and does not show any shift in the position of the cyclonic circulation at this height. Thus it is difficult to assess which of the factors is responsible for the slight shift at 1.5 km altitude. However, the good agreement in the circulation position at 3 km altitude indicates that the advection speed is likely to be well estimated, which suggests that the different resolutions of the dropsonde and Doppler measurements may be responsible for the obtained shift in the circulation position.

Noteworthy is the existence, in the upper-left part of the domain of Fig. 8(a), of an additional northerly flow associated with a clear-air region that is not sampled by the Doppler radar (see Fig. 8(b)). The location of this region within the secondary cyclone, using the satellite image of Fig. 3, indicates that this northerly flow is likely to correspond to a dry intrusion approaching the centre of the developing cyclone, since it is associated with an almost cloud-free zone, usually called the dry slot (Browning and Roberts 1996). This clearly illustrates that a complete description of the mesoscale dynamics of secondary cyclones must include descriptions of both the precipitation and clear-air regions (see also Joly et al. 1997).

The vertical structure of the dropsonde- and Doppler-derived airflows is now compared using vertical cross-sections. One of these cross-sections is given in Fig. 9, showing that both independent estimates of the airflow are characterized by quite similar structures; ascending motions are collocated with the region of stronger precipitation (as shown by the radar reflectivity of Fig. 9(b)) and descending motions on the right-side of the updraught. These downward motions reach the ground in both cases and deflect back to the left, which reinforces convergence in the lowest levels. Further comparisons of Figs. 9(a) and 9(b) indicate that the dropsonde- and Doppler-derived downward motions have the same magnitude, while the Doppler-derived updraught has a greater magnitude than the dropsonde-derived updraught. This shows that the mesoscale features are well-captured using the dropsonde sampling, but that the convective-scale to mesoscale updraught magnitudes, accurately recovered using the Doppler radar data, cannot be fully resolved using the dropsonde data. This result is not surprising, given the 70 km mean dropsonde spacing along the aircraft trajectories and the even larger spacing in the direction perpendicular to the straight-line flight tracks. Such identical characteristics are observed throughout the domain on any vertical cross-section. As explained previously, the vertical motion field deduced from dropsonde processing is expected to be more sensitive to an error in the advection speed than that deduced from Doppler radar. Thus the good agreement obtained seems to validate the advection speed used. In
addition, since this vertical wind component is roughly equal to the vertical component of the ageostrophic circulation, and since the vertical component of the ageostrophic circulation is likely to be more difficult to extract than its horizontal components, it seems reasonable to recover the 3-D ageostrophic circulation from this dropsonde-derived 3-D wind field.

Further comparison of the first-order derivatives of the dropsonde-derived and Doppler-derived horizontal wind fields confirms the previous result. Figures 9(c) and 9(d) show the dropsonde- and Doppler-derived estimates of the vertical component of the vorticity vector. Both vorticity fields exhibit roughly the same structure, with a low-level vorticity maximum of about 2 to 2.5 \times 10^{-4} \text{ s}^{-1}, and a vorticity tongue stretched along the frontal surface. The vorticity field is characterized by similar magnitudes, which is satisfactory since, as explained previously, the first-order derivatives are obtained in the Doppler radar case from measurements very close in space and time, which is not the case for the dropsonde-derived first-order derivatives. This comparison illustrates again the importance of combining the dropsonde and Doppler measurements. It is indeed clearly seen when comparing Figs. 9(c) and 9(d) that the clear-air vorticity structures missed in the upper levels by the Doppler radar in Fig. 9(d) are well-captured by the dropsondes in Fig. 9(c). A complete and accurate description of the mesoscale features associated with this secondary cyclone would thus be obtained, as expected, by combining the dropsonde (to access the clear-air parts of the cyclone) and Doppler (to get an accurate description of the vertical air motion magnitudes) measurements.
(b) Mesoscale dynamics of the secondary cyclone

In this paper it is not our intention to provide a detailed synoptic-scale interpretation of the IOP16 secondary cyclone. We rather concentrate on the mesoscale dynamics of the secondary cyclone sampled during the FASTEX IOP16 using aircraft dropsonde data, as these were specifically designed to define the mesoscale structure of the FASTEX secondary cyclones. However, we briefly recall in what follows the basic synoptic context in which this IOP16 secondary cyclone deepened. As explained in Joly et al. (1999), February was characterized by a strong zonal regime. The situation on 17 February was characterized by two parent low centres located over Iceland (L38*) and Greenland (L39).

The secondary low (L39A), a fast moving frontal wave sampled during IOP16, developed on the trailing cold front of L39. Low L39A developed near the south-west of the MSA and moved very rapidly to the north-east. It developed the typical cloud head of a deepening Atlantic storm depression, with associated heavy precipitation, pronounced dry intrusion, and strong winds (40 m s\(^{-1}\) at 900 hPa), as explained in Chaboureau and Thorpe (1999).

The satellite image of Fig. 3 exhibits a cloud cover typical of a mid-latitude frontal cyclone over the North Atlantic (see Browning and Roberts 1994 for a review), with: (i) a well-defined polar-front cloud band, associated with the warm conveyor belt (WCB, Harrold 1973), a synoptic-scale flow of air commonly observed, characterized by high \( \theta_e \), that advances polewards ahead of the cold front within mid-latitude frontal systems; (ii) an emerging cloud head (not fully developed) to the north-east of the polar-front cloud band, the appearance of which is a well-known symptom of cyclogenesis (e.g. Evans et al. 1994; Browning and Roberts 1994); and (iii) a dry slot forming between the polar-front cloud band and the emerging cloud head, characterized by few scattered clouds, indicative of a progressive intrusion of air from the upper troposphere and lower stratosphere towards the centre of a developing cyclone (called the 'dry intrusion', Reed and Danielsen 1959). The facts that the cloud head is not well-defined and that the dry slot is clearly not penetrating close to the centre of the developing cyclone indicate that the storm is not in a mature stage.

In addition to the horizontal wind vectors of Fig. 8(a), Fig. 10 shows the dropsonde-derived fields of pressure (Fig. 10(a)), equivalent potential temperature, \( \theta_e \) (Fig. 10(b)), relative humidity (Fig. 10(c)) and potential vorticity (Fig. 10(d)) at 1.5 km altitude. From the retrieved pressure field (Fig. 10(a)), the development of a part of the frontal wave into a secondary cyclone clearly appears, as shown by the well-defined pressure low (Fig. 10(a)) and associated cyclonic airflow circulation in the storm-relative frame (Fig. 8(a)). It is to be noted however that this secondary cyclone is not deep, since the cyclonic circulation is not visible in the ground-relative frame (see Fig. 6), which is generally the case for 'bomb-like' deepenings (e.g. Reed and Albright 1986; Lemaitre et al. 1999).

The \( \theta_e \) field (Fig. 10(b)) exhibits a well-defined baroclinic zone associated with the frontal surface, the lower part of the retrieval domain corresponding to the warm side of the front. Location of the horizontal airflow (Fig. 8(a)) with respect to the satellite image (Fig. 3) and the dropsonde-derived dynamic structures (Fig. 10(b)) shows that the south-westerly flow in the lower part of the domain is a high-\( \theta_e \) flow (Fig. 10(b)) collocated with the polar-front cloud band, which suggests that this flow is the mesoscale signature of the synoptic WCB. The westernmost part of this WCB is initially ascending and saturated on the left part of the domain. The shape of the high-\( \theta_e \) flow (greater than

* During FASTEX low-pressure systems were numbered in sequence, and those numbers are used here.
308 K) strongly suggests that it is the warmer air within the flow that deflects to the north while it ascends. The $\theta_e$ field at 3 km and 5 km altitudes (not shown) seems to validate this hypothesis, since the 308 K tempeature isocontours are progressively penetrating northwards on the right side of the cyclonic circulation, in association with the ascending part of the south-westerly flow. The southernmost part of this south-westerly flow is progressively descending (not shown) and cooling (see Fig. 10(b)) as it is located further east.

In the upper part of the domain, it appears clearly that the north-easterly flow is collocated with the developing cloud-head region (Fig. 3). This progressively ascending flow (as will be shown using air-parcel trajectory analysis) follows the frontal surface (Fig. 10(b)) and eventually wraps around the centre of the developing cyclone (on the left side of Fig. 8(a)). In contrast, the small part of a northerly flow sampled in the upper-left part of the domain is collocated with the almost cloud-free region. It is thus expected to be the mesoscale signature of the dry intrusion. This airflow is directed downwards (as shown by the cross-section of Fig. 11(b), perpendicular to the frontal surface)
and is much dryer than the north-easterly flow, as shown by the distinct signatures associated with these two airflows in terms of relative humidity (Fig. 10(c)). Those two characteristics are typical of dry intrusions before a cyclone reaches its mature stage. It is generally observed that this flow re-ascends as it approaches the secondary low when a cyclone is in its mature stage (as discussed in Young et al. (1987) and Browning and Roberts (1994)), which is effectively observed in our case, as will be shown in the cross-section through this flow (Fig. 12).

The $PV$ field, computed analytically from the dropsonde-derived 3-D wind and potential-temperature fields, exhibits a classic structure (Fig. 10(d)), with a positive $PV$ tongue (peak values of 1.5 PVU, where $1$ PVU $= 10^{-6}$ K kg$^{-1}$ m$^2$ s$^{-1}$) in the low levels associated with the cyclonic circulation, slightly stretched along the frontal surface.

We now turn to a detailed mesoscale description of the vertical structure of the secondary cyclone. For this purpose, two vertical cross-sections are used. The first vertical cross-section, discussed previously, is perpendicular to the frontal surface (Fig. 11), while the second one lies through the cyclonic circulation and the upper-level $PV$ and $PV_e$ anomalies, the structure of which will be shown and discussed hereafter (Fig. 12). The vertical cross-section through the frontal surface shows primarily the slope of the frontal surface (Fig. 11(a)), with a local maximum of $\theta_e$ below 2 km altitude, ahead (to the right) of the surface front. The component of the wind normal to the front (Fig. 11(b)) shows the lower part of the upper-level jet core (the jet core is not fully sampled by the dropsondes since they were launched from approximately 7 km, which
is below the jet core), located slightly to the warm-air side of the surface frontal zone, coincident with a region of strong cyclonic shear in the direction perpendicular to the cross-section. This component normal to the front extends downwards, as documented in particular by Shutts (1990), leading to winds of about 15 m s\(^{-1}\) in the warm air at 1.5 km altitude. The frontal zone, not fully resolved as a discontinuity by the dropsonde measurements, is also evident as the sloping layer of more pronounced vertical shear (Fig. 11(b)).

Wind vectors in the plane of the cross-front section (Fig. 11(b)) show that air is entering at the base of the main updraught from both sides, then turning upward and finally exiting the updraught above 4 km by deflecting continuously towards the cold air. This updraught is the ascending part of a direct ‘transverse’ ageostrophic circulation, with a classical updraught–downdraught structure associated with the warm- and cold-air sectors, respectively (Fig. 11(a)). Some upward motions are also found at mid to upper levels in the warm-air sector, in association with the lower part of the upper-level jet. It must be noted, however, that this cross-section only provides a 2-D view of the circulation, which implies that this ageostrophic circulation is not necessarily perpendicular to the frontal surface. Much of the tropospheric ascending branch of this ageostrophic circulation on the warm side of the front is close to saturation (see horizontal cross-section of relative humidity at 1.5 km altitude in Fig. 10(c)), until it reaches 4 km where it starts deflecting. In contrast, the descending air on the cold side of the front is much drier (Fig. 10(c)), which strongly suggests that this downward-directed and dry flow from the ground up to 3 km may be a developing dry intrusion. Unfortunately, due to the absence of dropsonde coverage within the almost cloud-free region of Fig. 3 (west of the upper-left border of the retrieval domain), the origin of this
cold airflow further west of the cross-section given in Fig. 11 cannot be assessed, and the dry intrusion cannot be identified with certainty.

The vertical cross-sections of PV and PV$_e$ are shown in Figs. 11(c) and 11(d), respectively. As predicted by models of frontogenesis and frequently observed (e.g. Thorpe and Clough 1991; Browning and Roberts 1994), the frontal zone is characterized by a positive PV anomaly in the lower troposphere, with peak values of 1.5 PVU between 1 and 2 km altitude. As discussed in Thorpe and Clough (1991), who found the same low-level PV signature using dropsonde data collected during the FRONTS 87 project, these maximum PV values, typically found at the tropopause level, are unambiguously not due to the descent of tropopause PV but rather due to the latent heat released in the saturated ascending air. The wind vectors in the plane of the cross-section of Fig. 11(c) confirm that, in our case, the low-level PV maximum is also collocated with upward air motions, which supports the observations of Thorpe and Clough (1991). An upper-level positive PV feature is also found above 6 km (Fig. 11(c)). This upper-level anomaly will be discussed in more detail by reference to the vertical cross-section through the cycloonic circulation (Fig. 12(c)).

The PV$_e$ cross-front section (Fig. 11(d)) is characterized by small values throughout the troposphere, which indicates that the atmosphere is roughly neutral to moist slantwise instability, except above 6 km and below 0.5 km where positive and negative PV$_e$ are found, respectively. Our estimate of the error on PV and PV$_e$ is about 0.4 to 0.6 PVU. As a result, the structures less than 0.5 PVU should not be interpreted as real. The low-level negative PV$_e$ anomaly has already been documented by Thorpe and Clough (1991) from dropsonde observations. Such a feature is generally attributed to the fluxes of moist entropy from the sea surface, which tend to generate negative PV$_e$. The assumption of a small and uniform PV$_e$ in frontal regions is generally used in models of moist frontogenesis (e.g. Thorpe and Emanuel 1985; Emanuel et al. 1987; Joly and Thorpe 1989). It has been shown in this context that the associated frontal structure was characterized by a local PV maximum near the surface front (due to latent-heat release), and that this low-level PV anomaly could be involved in the growth of frontal waves via secondary barotropic/baroclinic instabilities (Joly and Thorpe 1990, 1991). In addition, Emanuel (1985) and Thorpe and Emanuel (1985) both suggest that the neutrality to moist slantwise instability in saturated frontal regions (which corresponds to our case) significantly promotes frontogenesis, since the response to frontogenetic forcing is accordingly large.

As shown on the cross-section through the cyclonic circulation (Fig. 12), the PV field (Fig. 12(c)) exhibits a strong upper-level positive signature, with maximum values in excess of 4 PVU down to 4.5 km. This upper-level PV structure is found at lower altitudes than in Fig. 11(c), which suggests the existence of a tropopause fold in the clear-air region. This upper-level feature is also found in the PV$_e$ cross-section, (Fig. 12(d)), with a similar order of magnitude. In the other parts of the PV$_e$ cross-section, small positive values are obtained in the low- to mid-level troposphere, which again indicates that the atmosphere is roughly neutral to moist slantwise instability in these regions, except below 0.5 km where a negative signature is found, due to the fluxes of moist entropy from the sea surface (as discussed previously from Fig. 11(d)).

The upper-level PV and PV$_e$ anomaly seen in Figs. 11(c) and (b) and 12(c) and (d) is associated with very dry air (not shown). The dry feature with high PV in the upper-left part of this cross-section is commonly thought of as a signature of air coming from the lower stratosphere, induced by descent of the tropopause. In contrast, the right-hand part of the retrieval domain is a relatively moist region up to the 5.5 km altitude. This right-hand part corresponds to the warm side of the front, as shown by the slope and
location of the frontal surface in the \( \theta_e \) cross-section of Fig. 12(a). The component of the wind normal to the cross-section (Fig. 12(b)) shows the lower part of the upper-level jet core (as in Fig. 11(b)) associated with the upper-level PV anomaly, and the existence of a low-level jet on the warm-air side (see Fig. 12(a)). Wind vectors in the plane along the cross-section of Fig. 12(b) show a thermally indirect circulation, consistent with the existence of the previously discussed low-level and upper-level jets, with the ascending upper part of the circulation that crosses the lower part of the upper-level jet core. This vertical configuration of the jets and circulations can be compared with the coupling–uncoupling scheme proposed by Shapiro (1983). This comparison indicates that there is a slight coupling between the thermally indirect circulation of Fig. 12(b) and the ageostrophic circulation identified in Fig. 11(b), leading to a slightly favourable configuration for the development of ascending motions in the ascending branch of the ageostrophic circulation of Fig. 11(b). It must be noted that the observed coupling is not as developed as in the ideal case of Shapiro (1983), which may explain the relatively weak updraught obtained in our case.

(c) Air-parcel trajectory analysis

The air-parcel trajectory analysis is used in this section to give a better 3-D description of the internal mesoscale dynamics of the studied secondary cyclone. The basis of this air-trajectory analysis is given in SL. It uses the retrieved 3-D analytic wind field in the storm-relative moving frame to determine the successive positions of a given air parcel from a chosen initial position. The three trajectories most representative of the general mesoscale 3-D airflow circulation are given in Fig. 13, and discussed hereafter. These trajectories have been selected from systematic air-parcel trajectory retrievals so as to describe the internal dynamics of the studied cyclone.

The first representative trajectory starts at 1 km altitude (Fig. 13(a)), and is located on the northernmost part of the south-westerly flow, identified as the mesoscale WCB in section 4(b) (Fig. 8(a)). As anticipated in section 4(b), this flow is ascending to the domain top. It initially wraps around the cyclonic circulation of Fig. 8(a), as clearly shown by the projection of the 3-D trajectory on the ground, until it reaches a position of roughly \((x, y, z) = (0, -100, 2)\) in the retrieval domain. This behaviour of the selected air parcel is consistent with the structure of high-\( \theta_e \) at 1.5 km (\( \theta_e > 308 \) K, Fig. 10(b)), taking the form of a well-defined tongue of high-\( \theta_e \) at 1.5 km altitude interrupted from approximately the same position \((x, y) = (0, -100)\) and further east, resulting from the ascent of this flow. The trajectory analysis also shows that the warmer air within the south-westerly flow effectively deflects to the north while it ascends, as hypothesized in section 4(b), since an air parcel initially taken south of the parcel of Fig. 13(a) does not follow the same trajectory, but rather goes straight across the domain and slightly descends (not shown). The same air parcel (Fig. 13(a)) is then vertically accelerated and moves north-north-westward, along the flow that crosses the frontal surface at 3 km altitude (Fig. 8(c), roughly in the middle of the retrieval domain). Finally, instead of wrapping completely around the cyclonic circulation, it deflects back to the north in response to a strong northward flow at this location from 5 km altitude (not shown). This trajectory analysis thus indicates that the low-level vortex is not developed sufficiently to make the warm-air parcels circulate around the secondary cyclone centre, contrary to what is generally observed within deep frontal cyclones in their mature phase.

Two different trajectories representative of the northernmost and southernmost parts of the low-level north-easterly flow (Fig. 8(a)) are shown in Figs. 13(b) and 13(c), respectively. These trajectories are initially located in the lowest levels (approximately
Figure 13. Analysis of air-parcel trajectories representative of: (a) the south-westerly airflow of Fig. 8(a); (b) the northernmost part of the north-easterly airflow of Fig. 8(a); and (c) the southernmost part of the north-easterly airflow of Fig. 8(a). Initial and final positions of each parcel are denoted as I and F, respectively, and corresponding exact \((x, y, z)\) positions in the retrieval domain are given in the upper-left of each frame. Dotted lines represent projections of the trajectories to two dimensions.
500 m altitude) and roughly 60 km apart. The northernmost (Fig. 13(b)) and southernmost (Fig. 13(c)) parts of this flow both follow the frontal surface (see Fig. 10(b)) across the domain, as already discussed in section 4(b). The northernmost part (Fig. 13(b)) initially descends slightly and then progressively re-ascends up to 2 km as it approaches the secondary low centre. This slight ascent is consistent with the convergence obtained at 1.5 km altitude (Fig. 8(a)). Then, before it exits the retrieval domain, the northernmost part of the north-easterly flow (Fig. 13(b)) again descends slightly as it starts wrapping around the low-level vortex at 1.5 km. Unfortunately, due to the lack of dropsonde sampling west of the retrieval domain (already pointed out in section 4(b)), whether this air parcel keeps on wrapping around the low-level vortex and interacts with the south-westerly WCB cannot be inferred. In contrast, the southernmost part of the flow (Fig. 13(c)) experiences a completely different trajectory. It is indeed continuously ascending (up to 4.8 km when it exits the retrieval domain), wraps completely around the low- and mid-level vortex (from 1 km to 4 km altitude), finally exits the vortex and travels back to the south-west. The distinctly different behaviour of the two air parcels within the north-easterly flow may be due to the fact that the southernmost and northernmost parts of this flow are located on the warm and cold sides of the frontal surface, respectively, as confirmed by the \( \theta_e \) field of Fig. 10(b) which shows a 6 K temperature contrast between these two parts of the flow at 1.5 km altitude (298 and 304 K). Also noteworthy is the clear narrowing of the \( \theta_e \) gradients at 1.5 km in the convergence region of the two parts of the flow, roughly in the middle of the retrieval domain (Fig. 10(b)). It may then be postulated that the southernmost part of the north-easterly flow progressively overruns the colder northernmost part of this flow in response to the combined temperature contrast and low-level convergence of these flows.

5. SUMMARY AND FUTURE DIRECTIONS

In this paper, the mesoscale dynamics of a secondary cyclone sampled during the field phase of FASTEX is documented using airborne dropsondes released within the storm. A major interest in processing these dropsonde data is that the 3-D thermodynamic fields can also be retrieved from direct measurements which is not the case when Doppler radar data are used. In addition, dropsonde data are the only measurements capable of describing the clear-air regions of the secondary cyclones in the MSA that are not sampled by the airborne Doppler radars.

Before documenting the mesoscale dynamics of the IOP16 secondary cyclone, the capability of the method proposed to recover the mesoscale analytical 3-D fields of wind, pressure, temperature and relative humidity from the dropsonde measurements is evaluated in the present paper using simulated fields, airborne \textit{in situ} measurements, and airborne Doppler radar measurements in the same areas. All the simulations and comparisons with independent data sources indicate that the retrieval method is reasonably able to resolve the mesoscale structures of a frontal cyclone both in the regions of clear-air and precipitation. It is, however, found that the Doppler measurements are better at resolving the magnitude of the vertical air motion, due to the fact that the higher resolution of Doppler radar measurements allows description of smaller-scale structures that cannot be resolved using dropsonde sampling.

The secondary cyclone sampled during IOP16 of FASTEX was a fast moving frontal wave that developed on the trailing cold front of a parent low situated over Greenland (L39). This secondary low developed near the south-west of the MSA and moved very rapidly to the north-east. It developed the typical cloud head of a deepening Atlantic storm depression with associated heavy precipitation, pronounced dry intrusion, and
strong winds. Using the dropsonde-derived 3-D fields, the mesoscale flows involved in the organization of the secondary cyclone are described in detail. Interpretation of the PV and PV\textsubscript{e} fields shows the existence of a low-level PV anomaly associated with the frontal surface and a strong upper-level anomaly, while PV\textsubscript{e} remains small throughout the domain, except in the almost cloud-free region of the cyclone, where an upper-level anomaly is found, associated with a dry intrusion of stratospheric air. This indicates that the atmosphere is roughly neutral to moist slantwise instability, which, as shown by several authors (e.g. Joly and Thorpe 1990, 1991; Emanuel 1985; Thorpe and Emanuel 1985), significantly promotes frontogenesis.

The vertical configuration of the jets and circulations is compared with the coupling–uncoupling scheme proposed by Shapiro (1983), showing a slight coupling between the thermally indirect circulation and the ageostrophic circulation, leading in turn to a favourable configuration for the development of ascending motions in the ascending branch of the ageostrophic circulation. It must be noted, however, that the observed coupling is not as developed as in the ideal case of Shapiro (1983), which may explain the relatively weak upward vertical motions obtained in our case.

Further quantitative interpretation of the IOP16 secondary cyclone will be conducted in the near future. For this purpose, Doppler and dropsonde measurements will be combined, which should provide an accurate and complete description of the 3-D kinematic and thermodynamic fields both in regions of clear-air and precipitation. In particular, diagnostic quantities will be analysed so as to evaluate the current theoretical interpretations of secondary cyclogenesis, and to determine the multi-scale processes involved in the organization of the Atlantic secondary cyclones.

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