Role of the Low-Frequency Deformation Field on the Explosive Growth of Extratropical Cyclones at the Jet Exit. Part II: Baroclinic Critical Region

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ABSTRACT

Midlatitude cyclones tend to develop strongly in specific locations relative to the large-scale flow, such as jet-exit zones. Here, the approach developed in Part I that highlights the role of large-scale deformation in constraining the location of such events is continued. The atmospheric flow is decomposed into a high- and low-frequency part separating large and synoptic scales. A new low-frequency diagnostic has been introduced, called effective deformation $\Delta_{m}$. It is defined as $\sigma_{m}^{*} - \zeta_{m}^{*}$, where $\sigma_{m}$ is the low-frequency deformation magnitude and $\zeta_{m}$ is the low-frequency vorticity.

While Part I focused on large-scale conditions inducing an intermediate phase of barotropic growth, the present paper concentrates on other configurations that rather prevent this phase from happening. This large-scale circulation is characterized by the presence of a strong zonal upper-level jet and a lower-level jet that are meridionally quite far from each other over the Atlantic but close to one another in the eastern Atlantic region. As high-frequency disturbances are trapped by the effective deformation of the low-frequency jets, the increasing closeness of the two jets associated with that of the two effective deformation fields computed in the lower and upper levels defines a region called the baroclinic critical region where upper high-frequency disturbances and surface cyclones may strongly interact baroclinically. The increased baroclinic energy collection resulting from this constrained configuration change is outlined. An analysis of the explosive growth of the Christmas wind storms of 1999 and of mid-December 2004 provides different realizations of this configuration and associated mechanism.

1. Introduction

The investigation of the role of the large-scale deformation field on the evolution of synoptic disturbances is pursued in order to bring out new mechanisms that make extratropical cyclones strongly deepen at the jet exit. In Rivière and Joly (2006, hereafter Part I), specific regions in the deformation field are shown to be crucial for the development of these disturbances. They are called barotropic critical regions (denoted BtCR) as they correspond to very localized places (reduced often to a point) where the dilatation axes along the track of synoptic-scale disturbances change suddenly their orientation and where barotropic development tends to occur. Part I demonstrates that the presence of a BtCR region can be the source of cyclone development in the jet-exit region by illustrating it with the example of the Intensive Observation Period 17 (IOP17) of Fronts and Atlantic Storm-Track Experiment (FASTEX). However, some storms can also develop in the jet-exit region without the presence of a BtCR area. It is the case for the two extreme wind storms at the end of December 1999, called Lothar and Martin (hereafter denoted T1 and T2, respectively), which caused huge material damages and many human victims in western Europe (Ulbrich et al. 2001). The strongest storm that hit France since these two exceptional storms occurred on 17 December 2004 (hereafter denoted T3) and had a very similar development; it traveled across the Atlantic Ocean with moderate amplitude and strongly deepened at the jet exit but without the presence of a BtCR region. The question naturally addressed here, following the results of Part I, is: Does the large-scale deformation field play also a role in the location of these explosive growth stages despite the absence of a barotropic critical region?

The same methodology as in Part I is applied. It is based on a series of theoretical results from Rivière et al. (2003, 2004, hereafter RHK03 and RHK04, respectively) on perturbation growth in spatially and tempo-
rally complex quasigeostrophic flows. The use of the RHK results, as well as the decomposition of the atmospheric flow into a high- and low-frequency part, allows us to establish some useful low-frequency diagnostics to interpret the role of the large-scale deformation field on the path and growth of the high-frequency synoptic eddies. Furthermore, local energetics are used as the basic framework to confirm quantitatively our ideas.

Our methodology is summarized in section 2. Section 3 defines the concept of baroclinic critical region (BcCR) and illustrates it with the two shapes of zonal-like weather regimes of December 1999 and December 2004. Section 4 deals with the processes involved when a surface cyclone enters a baroclinic critical region. Section 5 discusses the relevance of preexisting upper-level high-frequency disturbances before the explosive growth stage of a surface cyclone. Concluding remarks are provided in section 6.

2. Methodology

The atmospheric flow is decomposed into a low-frequency part (denoted with subscript \( m \)) and a high-frequency one (denoted with primes). As in Part I, the low-frequency field is defined as the time mean over 8 days centered on the current date, and the high-frequency field is the subtraction of the total field from the low-frequency one.

a. Barotropic generation rate and effective deformation

Some results and notations of Part I are briefly recalled here. The barotropic generation rate can be written as the scalar product \( \mathbf{E} \cdot \mathbf{D}_m \), where \( \mathbf{E} = \left[ \frac{1}{2} (u''^2 - u'^2), -u' v' \right] \) and \( \mathbf{D}_m = \left[ \partial u_m / \partial x - \partial v_m / \partial y - v_m \tan \phi / a, \partial v_m / \partial x + \partial u_m / \partial y + u_m \tan \phi / a \right] \). Here \( u \) and \( v \) are the horizontal velocity components, \( \phi \) is latitude, \( a \) is the earth radius, \( \partial / \partial x = (a \cos \phi)^{-1} \partial / \partial \lambda \) and \( \partial / \partial y = a^{-1} \partial / \partial \phi \) are the horizontal derivatives, and \( \lambda \) is longitude. The barotropic generation rate is maximum (minimum) when the perturbation velocity vector aligns with the contraction (dilatation) axis. This leads to \( \mathbf{E} \cdot \mathbf{D}_m = +K'_e \sigma_m (-K'_e \sigma_m) \) where \( \sigma_m = |\mathbf{D}_m| \) is the deformation magnitude and \( K'_e = \frac{1}{2} (u'^2 - v'^2) \) is the perturbation kinetic energy.

However, the dilatation/contraction axes are not usually the equilibrium axes along which the perturbation orientation tends to align. In fact, two equilibrium axes exist whose exact formulation are recalled in Part I; one induces perturbation kinetic energy growth and is an unstable equilibrium, while the other induces perturbation decay and is stable. These orientations leads to the following approximate expression for the barotropic generation rate: \( \mathbf{E} \cdot \mathbf{D}_m = \pm K'_e \sigma_m \), where \( \sigma_m = \sigma_m^{\sigma_m} - \zeta'_m \) is the effective deformation and \( \zeta'_m = \partial u_m / \partial x - \partial v_m / \partial y - u_m \tan \phi / a \) is the rotation rate relative to the basis set \( \hat{i}, \hat{j} \) (\( \hat{i} \) and \( \hat{j} \) being the unit vectors directed eastward and northward). The sign of \( \Delta_m \) is crucial as it determines regions where equilibrium axes exist and where they do not. Positive values of \( \Delta_m \) correspond to regions where synoptic eddies can be strongly stretched because of the deformation action, and where barotropic processes can be very important whereas in regions of negative \( \Delta_m \) barotropic processes are quite weak. Part I has shown that \( \Delta_m \) is indeed a relevant tool to explain some barotropic processes in the atmosphere. It has been also emphasized from observational evidences that synoptic eddies tend to be trapped in areas of positive values of \( \Delta_m \). This second aspect will be particularly useful in the present study.

b. Instantaneous optimal baroclinic configuration

The baroclinic generation term that converts the available potential energy of the mean flow to eddy potential energy can be written as the scalar product \( \mathbf{F} \cdot \mathbf{B}_c \) (following the formalism of Cai and Mak 1990) where the two vectors \( \mathbf{F} \) and \( \mathbf{B}_c \) are defined by

\[
\mathbf{F} = \frac{1}{\sqrt{S}} \theta'(u' - u'),
\]

\[
\mathbf{B}_c = \left( -\frac{1}{\sqrt{S}} \frac{\partial \theta_m}{\partial y}, \frac{1}{\sqrt{S}} \frac{\partial \theta_m}{\partial x} \right).
\]

A simple way to interpret the baroclinic generation term is to divide it by eddy total energy \( T'_e = \frac{1}{2} (u'^2 + v'^2) + 1/(2S) \theta'^2 \). Here, \( \theta \) is the potential temperature and \( S \) is the static stability. This leads to the following expression:

\[
\frac{\mathbf{F} \cdot \mathbf{B}_c}{T'_e} = |\mathbf{B}_c| \cos(\mathbf{F}, \mathbf{B}_c) = |\mathbf{B}_c| \left| \frac{\mathbf{F}}{T'_e} \right| \cos(\mathbf{F}, \mathbf{B}_c),
\]

where

\[
\frac{|\mathbf{F}|}{T'_e} = \frac{\sqrt{ \frac{1}{5} \theta'^2 (u'^2 + v'^2) }}{\frac{1}{2} (u'^2 + v'^2) + \frac{1}{2S} \theta'^2 },
\]

and \( \cos(\mathbf{F}, \mathbf{B}_c) \) is the cosine of \( \mathbf{F} \) and \( \mathbf{B}_c \). The term \( \mathbf{F} \cdot \mathbf{B}_c / T'_e \) is called the exponential baroclinic generation rate as it appears in the equation governing the total energy exponential growth rate. It is the product of the baroclinicity \( |\mathbf{B}_c| \) with a term, hereafter called configuration term and denoted conf, corresponding to the
eddy spatial configuration relative to the baroclinicity vector orientation. The configuration term is itself the product of two terms, $|F|/T'$ and $\cos(F, B_c)$, which are related to two well-known different notions of instantaneous optimal baroclinic configuration; $\cos(F, B_c)$ is maximum when the two vectors $F$ and $B_c$ are colinear. If we consider the interaction between an upper disturbance and a surface cyclone, the $F$ vector is mainly oriented along the axis formed by the two disturbances. If $F$ and $B_c$ are not colinear, it means therefore that the upper-level disturbance does not align with the surface cyclone along the baroclinicity vector and the high-frequency perturbations do not extract optimally the available potential energy of the mean flow. Such a situation is thematically presented in Fig. 1a. The second quantity of the configuration term, $|F|/T'$, is related to the tilt with height of the high-frequency perturbation isolines. Indeed, following Eq. (4), $|F|/T'$ is maximum when $u'^2 + v'^2 = (1/S)\theta^2$, that is, when kinetic energy equals to potential energy; this equality is not satisfied if the high-frequency isolines are strongly tilted, as $u'^2 + v'^2 \ll (1/S)\theta^2$ in this case (see Fig. 1b), or if they are almost vertical, as it implies $u'^2 + v'^2 \gg (1/S)\theta^2$ (see Fig. 1c). The equality $u'^2 + v'^2 = (1/S)\theta^2$ defines therefore a specific vertical slope for the high-frequency geopotential isolines that can be found by approximating $u', v'$ with the horizontal derivatives of the geopotential and $\theta'$ with the vertical one. The condition $|F|/T' = 1$ is equivalent to the phase quadrature condition between the upper-level wave and lower-level one as discussed in simple idealized models (see e.g., Hoskins et al. 1985; Warrenfeltz and Elsberry 1989; Davies and Bishop 1994). The configuration that optimally extracts energy from the mean flow is shown in Fig. 1d. Let us emphasize that the optimal baroclinic configurations discussed previously are considered only in the instantaneous sense of the term and no notion of time duration intervenes.

c. Dataset

The data for December 1999 are provided by re-analysis dataset at 0000, 0600, 1200, 1800 UTC of the Météo-France operational four-dimensional variational data assimilation (4DVAR) Action de Recherches Petite Echelle Grande Echelle (ARPEGE) system. We also use forecast datasets of the same period but at 0300, 0900, 1500, and 2100 UTC; 3 hourly steps are useful here to analyze the development of the two storms as their spatial structure and growth change very rapidly with time. Note that this dataset has the noticeable advantage over other ones to be the only model to consistently forecast the trajectory of the first storm over a wide set of forecast ranges. Few complementary results are provided by analysis dataset of the same operational system but for the period of mid-December 2004.

3. Baroclinic critical regions

Figure 2 is a sketch defining the main features of a baroclinic critical region. It appears in a kind of zonal
A weather regime composed of a lower- and an upper-level jet that are well separated in regions where their speed are maximum (left-hand side of the figure) but the separation distance is then decreasing farther eastward where the wind speeds of the jets are also decreasing (right-hand side of the figure). This property can be deduced from semigeostrophic theory, the stronger the wind speed of the jets, the larger the distance between the two jets. Its main distinctive feature is that regions where the effective deformation $\Delta_m > 0$ remain on the cyclonic side of the jet stream both upstream and downstream of the jet exit that lies at the core of the region. However, the reducing speed of the jet as one nears the exit implies, in a balanced atmosphere, that the vertical slope of a number of properties such as isentropes or absolute momentum surfaces increases: the low-level jet closes on the location of the upper-level one. This is the main feature of a baroclinic critical region. A cyclone moving through it, shown by the four-step sequence of heavy solid contours of vorticity or geopotential, as in previous situations, also has to remain in areas where $\Delta_m > 0$. This implies that, as the system nears the exit area, its low-level component tends to close on the upper-level one and to align with it in the direction of the large-scale baroclinicity (step 2 and step 3). At step 3, provided the upper-level component remains upstream, the vertical structure is in line with $\mathbf{B}$, so that it is optimal at least for that component of the baroclinic source term, and growth is increased, possibly significantly, depending on the other aspects.

Fig. 2. Schematic representing a BeCR with the same drawing conventions as those shown in Part I. The large-scale flow making a baroclinic critical region is such that areas where $\Delta_m > 0$ remain on the cyclonic side of the jet stream both upstream and downstream of the jet exit that lies at the core of the region. However, the reducing speed of the jet as one nears the exit implies, in a balanced atmosphere, that the vertical slope of a number of properties such as isentropes or absolute momentum surfaces increases: the low-level jet closes on the location of the upper-level one. This is the main feature of a baroclinic critical region. A cyclone moving through it, shown by the four-step sequence of heavy solid contours of vorticity or geopotential, as in previous situations, also has to remain in areas where $\Delta_m > 0$. This implies that, as the system nears the exit area, its low-level component tends to close on the upper-level one and to align with it in the direction of the large-scale baroclinicity (step 2 and step 3). At step 3, provided the upper-level component remains upstream, the vertical structure is in line with $\mathbf{B}$, so that it is optimal at least for that component of the baroclinic source term, and growth is increased, possibly significantly, depending on the other aspects.
pict the case where the upper and lower disturbances are located around the jet-core region. As the two levels' perturbations are forced to be in regions of positive $\Delta_m$, the upper-level disturbance is necessarily located northwest of the surface cyclone. The $\mathbf{F}$ vector has a NW–SE orientation but the baroclinicity vector $\mathbf{B}_c$ is almost zonal leading to a large acute angle between $\mathbf{F}$ and $\mathbf{B}_c$. Moreover, the two disturbances are quite far from each other, suggesting that the high-frequency isolines are strongly tilted. Therefore, none of the two notions of optimal instantaneous configuration described in section 2b is satisfied in this jet-core region.

By contrast, when the two disturbances reach the jet-exit region (steps 3 and 4), areas of positive $\Delta_m$ at the two levels overlap; there is no more constraint to refrain the two disturbances from getting the optimal baroclinic configuration. The composite structure they make up together can be first aligned with the baroclinicity vector $\mathbf{B}_c$ and second sufficiently close to each other to have the right tilt with height while satisfying the constraint of location in a region of positive $\Delta_m$. Although the magnitude of the baroclinicity vector $|\mathbf{B}_c|$ is decreasing in the jet-exit region, a favorable baroclinic configuration in the jet-exit region can compensate for this to the extent that the baroclinic interaction in the jet-exit region is stronger than in the jet-core one. To conclude, as high-frequency disturbances in the upper and lower levels are both forced to evolve along the corridors determined by regions of positive $\Delta_m$, a strong interaction is possible between the two types of disturbances only in the area where regions of positive $\Delta_m$ in the upper and lower levels converge, that is, in the baroclinic critical region. Two examples of such a large-scale circulation configuration are now provided.

A detailed analysis of the low-frequency upper-level jet (350 hPa) at the end of December 1999 has been already performed in Part I. In particular, a barotropic critical region has been identified in the jet-entrance region around the point $(43^\circ N, 57^\circ W)$. It is visible in the left-hand side of Fig. 3a as this point separates two regions of positive $\Delta_m$ where dilatation axes are perpendicular upstream and downstream of it. But from the jet core to its end, regions of large positive $\Delta_m$ are all located on the northern side of the jet and no barotropic critical region is present. At low levels (850 hPa), a narrow jet appears clearly between $50^\circ$ and $20^\circ$ W and extends toward $0^\circ$. As for the upper levels, regions of positive $\Delta_m$ in lower ones are all located on the cyclonic side of the jet. Between $40^\circ$ and $20^\circ$ W, large values of the effective deformation fields form two narrow corridors north of their respective jets suggesting very localized and very strong shears at that place. These two corridors are very far from each other in regions where the two jets reach their maximum speed with a separation distance of almost $10^\circ$ in latitude. But this separa-
ition is decreasing eastward as the wind speeds of the jets are decreasing and the two regions of positive $\Delta_m$ converge toward each other around 20°W, which defines a baroclinic critical region at that longitude. Figure 3b presents the high-frequency relative vorticity densities computed at 350 and at 850 hPa between 24 and 28 December (an exact definition is provided in appendix A of Part I). There is a clear latitudinal shift of the upper-level disturbances density related to that of the surface cyclones around 40°W but the two densities begin to be superimposed around the BeCR region. Figures 3c,d show that at both levels the high-frequency densities are located in regions of positive values of $\Delta_m$ between 50° and 20°W on the north side of the jet confirming a result of Part I that high-frequency eddies’ location is well captured by the positives values of the effective deformation. The high-frequency disturbances at 850 hPa tend to propagate along the southwest–northeast (SW–NE)-oriented corridor formed by the effective deformation at that level whereas those at 350 hPa follow the zonal orientation of the upper-level effective deformation. To summarize, Fig. 3 reveals the presence of a BeCR region in the large-scale circulation at the end of December 1999 and the convergence of the two high-frequency densities in this region suggests that upper disturbances strongly interact with surface cyclones around it.

The three-dimensional structure of the low-frequency jet of mid-December 2004 is very similar to the previous one. Indeed, at upper levels (Fig. 4a), a clear BtCR region appears in the jet-entrance region around the point (45°N, 60°W) but such a region does not exist from the jet core to its exit as in the December 1999 case. The low-level jet of December 2004 (Fig. 4a) is located far south of the upper-level jet and has a region of positive $\Delta_m$ very confined meridionally on its northern side that reaches west coasts of France and England. There is a clear superimposition of the $\Delta_m$ fields computed in the upper and lower levels around Ireland and west coast of England, which defines a BeCR region. The trajectory of the surface cyclone leading to the strong windstorm that hit France on 17 December 2004 is shown in Fig. 4b. It follows exactly the corridor formed by regions of positive $\Delta_m$ at low levels and its explosive phase of growth occurs between 0000 and 1800 UTC 17 Feb 2004 in the baroclinic critical region.

In the following sections, the aim is to provide actual realizations of the mechanism outlined here, namely that the wind storms explosive growth in the jet-exit region can be explained by a strong transient baroclinic interaction due to a favorable baroclinic configuration, which is allowed in a BeCR region and not upstream of it. A fully quantitative demonstration is based on the detailed description of the explosive growth of T2.

4. Efficient baroclinic configuration in a BeCR region

In what follows, the notations Day1, Day2, Day3 refer to 25, 26, and 27 December 1999, respectively.

a. Detailed description of T2

The time evolution of T2 is represented in Fig. 5 by the high-frequency relative vorticity field at 850 hPa (dashed black contours). The trajectory of the storm is parallel to an axis oriented SW–NE, which is linked to the low-level jet axis, and its structure is located in regions of positive values of $\Delta_m$ at low levels as expected from the results of the previous section. The effect of the deformation field on T2 is well visible as it is strongly elongated at 1200 UTC Day2 (Fig. 5a). In the appendix, the storm is shown to be more stretched along the stable axis deduced from the theory described in section 2a than along the dilatation axis, a result that underlines the consistency of our approach. Here, T2 interacts with a well-defined preexisting upper-level disturbance during its whole life cycle (Hello and Arbogast 2004), which is visible in Figs. 5a–c (see solid
black contours). From 1200 UTC Day2 to 0000 UTC Day3 (Figs. 5a,b), the upper disturbance crosses the jet from the south to the north. Part I provides a rationale for this crossing: the perturbation loses energy barotropically on the anticyclonic side of the jet whereas it is regenerated on the cyclonic side. The result is that the upper-level disturbance is almost located north of the upper-level jet at 0000 UTC Day3 and most of its structure is northwest of the surface cyclone at that time. This spatial configuration is similar to step 2 of Fig. 2, suggesting a nonoptimal baroclinic configuration. Twelve hours later (Fig. 5c), the surface cyclone has evolved more northward and is now at the same latitude as the upper disturbance and the two disturbances are closer to each other. This configuration is similar to step 3 of Fig. 2 suggesting an optimal baroclinic configuration, which is confirmed in the more quantitative following section.

Figure 6a represents the time evolution of volume integrals of the baroclinic generation rate \(T_f^B_{\text{conf}}\), the low-frequency baroclinicity \(|B_c|\), the eddy total energy \(T'_{\text{en}}\), and the exponential generation rate \(|B_c|_{\text{conf}}\) [see Eq. (3) for their definitions]. The baroclinic generation rate (thick solid line with diamonds) has a well-defined peak at 1500 UTC Day3. This peak is not at all due to a baroclinicity maximum as it occurs when baroclinicity is decreasing (dash–dotted line). This result is not astonishing as the storm strongly intensifies in the jet-exit region, and thus far away from the baroclinicity maximum. But the exponential baroclinic generation rate \(|B_c|_{\text{conf}}\) (dashed line in Fig. 6a) is maximum exactly at the same time as the overall baroclinic generation rate showing that the latter is strongly dependent on the configuration term conf, that is, on the baroclinic configuration of the disturbances related to the baroclinicity vector. The fact that the baroclinic generation rate keeps changing, increasing strongly first and then immediately decreasing just as strongly reflects the rapid change of configuration of the perturbations and rules out the idea that some phase locking take place during rapid cyclogenesis.

The kinetic energy growth of the surface cyclone is now analyzed more precisely. The time evolution of the high-frequency kinetic energy maximum computed at low levels around T2 (Fig. 6b) is compared to that of the barotropic generation rate and the baroclinic conversion rate (Fig. 6c). The last term is the baroclinic term that converts eddy potential energy to eddy kinetic energy. The kinetic energy maximum at low levels is analyzed instead of an average, to better understand the peak of velocity of this windstorm responsible for huge damages in western Europe. During the period of strong kinetic energy growth, the baroclinic conversion rate is very strong and has a well-defined peak at 1500 UTC Day3, which corresponds exactly to the peak of the baroclinic generation rate shown in Fig. 6a. These peaks of the baroclinic conversion rates slightly precede the peak of eddy kinetic energy maximum and take place when the slope of the eddy kinetic energy curve is the strongest. The barotropic generation rate by contrast is significantly smaller than the baroclinic conversion rate and does not play a role in the explosive growth of the two storms. The fact that barotropic processes are negligible in the jet-exit region was expected as no barotropic critical region was present from the jet core to the jet exit. The barotropic generation rate only has a large signal in the upper troposphere between 0000 UTC Day2 and 0000 UTC Day3 when the upper-level disturbance associated with T2 crosses the jet as shown in Part I. This signal is not present however in
the curve of Fig. 6c as the latter pertains to low levels only (between 500 and 900 hPa).

Concerning the redistribution terms of energy (not shown here), the term related to the horizontal ageostrophic geopotential fluxes is negative, on average, during the kinetic energy growth of each storm and acts therefore to disperse energy.

Figure 7 describes more precisely the baroclinic configuration of T2 in order to compare it with the schematic picture of Fig. 2. At 0000 UTC Day3 (Fig. 7a), the $\mathbf{F}$ vector is oriented NW–SE as it was predicted in Fig. 2 (step 2), which confirms that there is baroclinic interaction with the upper-level high-frequency structure located northwest of T2 around the point (50°N, 30°W).

As $\mathbf{B}_c$ is almost zonal, the cosine of $\mathbf{F}$ and $\mathbf{B}_c$ is positive but not close to 1, proving that baroclinic interaction is present but not optimal. Figure 7c is a vertical cross section of the high-frequency geopotential isolines at the same date, plotted along the $\mathbf{F}$ vector, and not along $\mathbf{B}_c$ as it is done usually, as $\mathbf{F}$ represents the main direction of the baroclinic interaction. On this cross section, one can see T2, the upper-level structure upstream of it, and a clear northwestward tilt with height of the high-frequency geopotential isolines confirming the presence of baroclinic interaction. At 1500 UTC Day3 (Fig. 7b), T2 is already in the jet-exit region. Despite the decrease of the baroclinicity magnitude $|\mathbf{B}_c|$ (see the decrease of the magnitude of the gray arrow), the baroclinic generation rate is three times larger at 1500 UTC than at 0000 UTC Day3 because the $\mathbf{F}$ vector is now along the baroclinicity vector $\mathbf{B}_c$. The geopotential isolines have now a pure westward tilt with height, the baroclinic configuration is optimal and similar to step 3 of Fig. 2.

b. Brief description of T1 and T3

The strong deepening of T1 at the jet exit is quite similar to that of T2 in several aspects. Its trajectory follows almost the same path as T2; that is, it evolves along regions of positive $\Delta_m$ located north of the low-frequency low-level jet and its explosive growth occurs about the BeCR region. Figure 8 describes a part of the evolution of T1. At 1200 UTC Day1 (Fig. 8a), its structure is in a region of positive $\Delta_m$ and an upper-level high-frequency structure is located 5° north of it (see solid black contours). The latitudinal shift of the upper-level effective deformation related to that computed at low levels is responsible for the latitudinal shift of the two-level high-frequency disturbances but as the baroclinicity is zonal (see arrows), strong baroclinic interaction between the two high-frequency structures cannot happen. Because of the SW–NE orientation of the low-frequency jet, T1 evolves more and more northward,
reaches the latitude of the upper-level high-frequency structure at 0000 UTC Day2 (Fig. 8c), and begins its explosive growth. Although the two maxima of vorticity at 350 and 850 hPa at that time are not aligned with the baroclinicity vector, they are close to each other, which suggests an improvement of the second aspect of the baroclinic configuration.

The strong development of T3 (Fig. 9) is quite similar to the previous scenario. At 1800 UTC 16 December 2004 (Fig. 9a), the low-level cyclone which has a small-scale structure as T1 is in region of positive $\Delta m$ and high-frequency upper-level structures are visible north of it. These locations can be still explained by the latitudinal shift of the two-level deformation fields. It refrains the two disturbances from getting an alignment along the zonal orientation of the baroclinicity vector. However as the two deformation fields are superimposed in the jet-exit region, the two disturbances become closer at 0600 UTC 17 February (Fig. 9b) and their associated maxima of vorticity align more and more with the baroclinicity vector (see Figs. 9c,d) as the properties of the effective deformation fields allow it.

Note that the gray arrows corresponding to the baroclinicity vector are twice as large in the position of the storm at 1800 UTC 16 December (Fig. 9a) before its explosive growth stage than at 1200 UTC 17 December (Fig. 9d) during it. The figure suggests here again an improvement of the baroclinic configuration in the jet-exit region far downstream of the baroclinicity maximum.

A strong transient baroclinic interaction related to a transient optimal baroclinic configuration of the high-frequency disturbances is responsible for the strong intensification of T2 in the jet-exit region. The developments of T1 and T3 in the jet-exit region as discussed from the last figures suggest an optimization of the baroclinic configuration about a BeCR region that may explain the location of their explosive growth at the jet exit.

Diabatic terms probably play a role in explaining the final intensity reached by these phenomena. However, unlike the reading that can be made of Wernli et al. (2002, hereafter W02) where diabatic processes appear to explain the whole occurrence of such storms, the
present view is that they “only” amplify the consequences of the dynamical interaction presented in the present study. Notice, in particular, that the developments occur not only away from the maximum baroclinic area, but also away from the warmer and moister area, away from the smaller static stability area, all conditions that would be essential in locating a purely diabatically induced development.

Fig. 8. Time evolution of T1 with the same definitions as in Fig. 5. The baroclinicity vector (gray arrows) is added. (a) 1200 UTC Day1, (b) 1800 UTC Day1, (c) 0000 UTC Day2, and (d) 0600 UTC Day2.

Fig. 9. Time evolution of T3 with the same definitions as in Fig. 8 but the upper level is here 300 hPa. (a) 1800 UTC 16 Dec, (b) 0000 UTC 17 Dec, (c) 0600 UTC 17 Dec, and (d) 1200 UTC 17 Dec 2004.
5. Discussion

Despite the similarity of the two Christmas wind storms of 1999 in terms of their trajectory and the location of their explosive growth stage, some differences in terms of the scenario exist also that are important in order to discuss the relevance of the Sutcliffe–Petterssen scheme or type-B development (Petterssen and Smebye 1971). The proposed mechanism to explain the location of the explosive phase of the two storms, as summarized in the previous paragraphs, would imply that there exists preexisting upper-level disturbances interacting with the surface cyclones T1 and T2. It is clear that T2 interacts with the same well-defined upper disturbance during its whole life cycle as Hello and Arbogast (2004) have shown and as it has been emphasized in the present paper, but the interaction of T1 with a well-defined preexisting upper disturbance seems to be more questionable. W02 have pointed out that the development of T1 is characterized by a shallow low-level cyclone of moderate intensity, which intensifies strongly as it crosses the jet-stream axis and no preexisting well-defined upper-level disturbance seems to interact with T1 during its whole life cycle. W02 explains the appearance of an upper-level disturbance during the crossing of the jet by a bottom-up development, which is an equivalent of Fig. 21 of Hoskins et al. (1985); the circulation induced by the low-level potential vorticity (PV) anomaly in the upper troposphere advects PV southward because of the strong PV gradient around the jet maximum and creates an upper-level PV anomaly just upstream of the low-level vortex. It seems a priori to contradict our results as we have shown that there exists already just north of the jet axis a high-frequency upper-level structure before the explosive growth phase of T1 (see, e.g., Fig. 8b) even though it is not a well-defined coherent disturbance. From our high- and low-frequency decomposition, it is difficult to state whether the upper-level disturbance just upstream of the surface cyclone during the explosive growth phase of T1 (Fig. 8c) is the same as the high-frequency upper-level disturbance located 6 h before northwest of the surface cyclone (Fig. 8b), or, as the argument of W02 suggests, whether it is created by the advection of low-frequency PV southward by the circulation induced by the low-level high-frequency PV anomaly. It seems that both the preexisting high-frequency upper-level structure, and the mechanism described by W02 can be responsible for the presence of the well-defined upper-level disturbance in Fig. 8c. Let us emphasize however that our argument on the location of the explosive phase of growth is relevant also when the mechanism depicted by W02 is predominant. Indeed, if it was the case, the low-level cyclone would be forced by the effective deformation of the low-frequency jet to follow a SW–NE trajectory. It would only become able to stir a large upper-level PV gradient in the jet-exit region and not before and therefore to start an explosive growth only at this moment. To conclude, the upper disturbance that interacts with T2 is clearly visible and supports the classical type-B mechanism whereas the upper-level disturbance associated with T1 is more difficult to follow as a coherent structure evolving in space and time. An upper-level high-frequency disturbance has been identified for the storm of December 2004 (Fig. 9) before its explosive growth but as for T1, its amplitude is quite moderate and it does not appear to be a coherent structure either. Further studies are necessary to investigate the important role or not of these preexisting high-frequency structures prior to the explosive growth stage, beginning by improving their definition and identification in real fields.

6. Concluding remarks

Midlatitude cyclones evolving in the northeast Atlantic grow in the jet-exit region and their strongest deepening rates tend to occur downstream of the baroclinicity maximum (see, e.g., Fig. 15 of Wang and Rogers 2001). It was in particular the case of the two Christmas wind storms of 1999 as well as that of mid-December 2004. These results suggest that large baroclinicity is a necessary condition for the development of strong wind storms but cannot give us any information on the location of their explosive growth stage. The aim of the paper was to show how the large-scale deformation field and the large-scale vorticity compete to finally determine the location of synoptic disturbances and force them to interact baroclinically in specific regions. The new effective deformation \( \Delta_m \) embodies this competition between deformation and rotational effects. The key fact brought out by the present approach is that at each level, the high-frequency disturbances are trapped in regions where the effective deformation \( \Delta_m \) is positive.

The two shapes of zonal-like weather regimes that lead to the development of the last strongest wind storms in France (Christmas 1999 and mid-December 2004) are quite similar. Both are characterized by exceptionally strong jets exhibiting not only strong baroclinicity but also narrow and strong horizontal shears. Regions where the effective deformation \( \Delta_m \) is positive are therefore very localized and these regions computed in the upper and lower levels form two narrow corridors north of their respective low-frequency jets.
that converge over the eastern Atlantic region, in an area called in the text baroclinic critical region.

The constraints of location imposed by $\Delta_m$ added to the convergence of the two low-frequency jets in the baroclinic critical region show that the surface cyclone and the upper-disturbances, when both exist, can strongly baroclinically interact only in the baroclinic critical region. Indeed, the argument is the following; suppose that an upper disturbance interacts baroclinically with a surface cyclone in the jet-core region, this baroclinic interaction cannot be optimal due to the constraints of location of each disturbance in regions of positive $\Delta_m$ because first, the upper disturbance cannot align with the surface cyclone along the baroclinicity vector and second, the two disturbances are quite far from each other. By contrast, in the jet-exit region, such constraints are not present, the upper disturbance can be sufficiently close to the surface cyclone to get the appropriate tilt with height and the alignment with the baroclinicity vector can be satisfied too. Let us note that this qualitative argument based on the location of the effective deformation does not suggest that there will be necessarily an upper disturbance each time that will interact with the surface cyclone, nor that the deformation field helps the two disturbances to stay in phase. It does not say either that the two notions of baroclinic configuration will be necessarily optimized in the jet-exit region, but allows to say that if a surface cyclone has an explosive growth in such a weather regime, it can only occur in the baroclinic critical region.

A detailed description of T2 proves that this mechanism does exist in the real atmosphere. The favorable phase of baroclinic configuration between the high-frequency disturbances in the upper and lower levels did not last (=12 h) and is therefore very far from a normal mode behavior and in this sense has also to be distinguished from a phase locking configuration as that described in Hoskins et al. (1985).

The results of the present paper go in the same direction as those of James (1987) who has shown in horizontally sheared flows that the barotropic component of the basic flow confines perturbations to be located in specific regions that do not correspond to the most baroclinic part of the basic flow and baroclinic interaction is thus reduced by the presence of the shears. In the same manner, it has been proved here that the confinement of synoptic eddies in regions where $\Delta_m$ is positive leads to baroclinic interaction far away from the baroclinicity maximum. The advantage of the low-frequency effective deformation $\Delta_m$ is that it can be applied to any observed low-frequency jet.

The theoretical study of Hare and James (2001) has already shown that the constraints imposed by the basic state properties can lead to a meridional displacement of the upper and lower disturbances and to a reduction in the efficiency of the baroclinic interaction. But by contrast with this study that focuses on the effect of the ageostrophic meridional circulation on disturbances location, our work shows that this meridional displacement is due to a combination of both a strong baroclinicity and a strong horizontal deformation.

The “crossing of the jet” from the warm to the cold air side in region of diffluence seems to be the preferred place for strong deepening rates of northeastern Atlantic cyclones. The aim of the present paper was to investigate further dynamical processes leading to strong cyclone growth during the crossing of the jet stream in its exit region by making explicit scale interaction between disturbances, and their slowly evolving environments. Part I and the present paper exhibit two different realizations of crossing of the jet. In Part I, it is shown that during the explosive phase of growth, the upper-level disturbance and the surface cyclone, respectively, cross the upper- and lower-level jets from the south to the north and barotropic generation rates play a significant role because of the rapid change of the dilatation axes that characterizes the first configuration. This in turn creates or regenerates a baroclinic tilt with height which can take the growth over. In this paper, both the upper- and lower-level disturbances stay on the northern side of their corresponding jets in the eastern Atlantic. In the latter case, the surface cyclone crosses the upper-level jet, but not the lower-level one, and barotropic processes are thus not important during this crossing. A change in the constraints of location imposed by the deformation field along the jet allows the development of a more optimal baroclinic configuration between the upper- and lower-level disturbances in the jet-exit region, which results in the growth.

Finally, it must be emphasized that the low-frequency effective deformation $\Delta_m$ has an important predictability potential to help localize regions where wind storms will reach their maximum amplitude. In the jet-exit region, $\Delta_m$ is useful to anticipate regions where storms will have to go through. For example, large values of $\Delta_m$ are located north of England for the jet of mid February 1997 (see Fig. 2 of Part I), where IOP17 have gained its maximum amplitude. The same diagnostic for December 1999 is centered over France and Germany (see Figs. 5 or 8) exactly where T1 and T2 have reached their maximum amplitude and have made huge damages. Two zonal jets apparently differing only in terms of their amplitudes have thus subtle but distinguishable structure differences in their horizontal inhomogeneities that may lead to very different behaviors of midlatitude cyclones life cycles and especially in
their location. The fact that these differences depend solely on a large-scale, slowly evolving property is of major interest both in terms of understanding cyclogenesis as a scale interaction process and for medium-range prediction of wind storm risk assessment.

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APPENDIX

Alignment Properties

Figures A1a,b describe the alignment properties of the two surface cyclones associated with T1 and T2 during their evolution north of the low-level jet. Following the theory of RHK and Part I, the perturbation is not exactly stretched along the dilatation axis [defined by the angle $0.5(\pi/2 - 2\phi_m)$, where $(iD_m) = \pi/2 - 2\phi_m$] but along a stable axis whose angle with respect to the x-axis is $0.5[\arccos(\pi/2 - 2\phi_m)]$. The two different axes are respectively plotted in Figs. A1a,b with gray and black arrows. At 0600 UTC Day1 the surface cyclone associated with T1 is slightly stretched in the zonal direction, almost in the same direction as the stable axis whereas the dilatation axis has a pronounced NW–SE tilt. At 1200 UTC Day2 the surface cyclone associated with T2 is strongly stretched zonally also and even if the stable axis is not completely zonal, the elongation of T2 is closer to the stable axis than to the dilatation axis. This result shows the consistency of our approach and the nonnegligible role played by the low-frequency vorticity on the orientation of the high-frequency disturbances. Generally, the name dilatation axis is not very appropriate as the structure of a perturbation is not dilated along this axis. As shown in RHK03 and RHK04, it is only in very specific cases when the rotational component of the deformation field is zero, as for example in the pure strain case, that it may occur.

REFERENCES


