FASTEX IOP 18: A very deep tropopause fold. II: Quasi-geostrophic omega diagnoses

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SUMMARY

A set of two companion papers is dedicated to the documentation of the life cycle of a very deep tropopause fold (820 hPa) during the Intensive Observing Period 18 of the Fronts and Atlantic Storm-Track EXperiment (FASTEX). In this second part, diagnoses of vertical motion using a Q-vector partitioning in the natural coordinate system that follows the geostrophic wind are analysed. The partitioning allows the evaluation of vertical motions associated with forcing mechanisms such as confluence and diffluence, thermal advection by the horizontal geostrophic shear (shear advection) and curvature of the flow. The synoptic situation involves the formation of an intense upper-level jet streak when an Arctic trough and a southern ridge move into phase with each other. Results assessed in the course of the tropopause-fold life cycle show that subsiding vertical motions associated with each of the forcing mechanisms (confluence, shear advection and curvature) overlap on the cyclonic-shear side of the entrance region of the jet streak. It is shown that an additional effect of shear advection over the confluence, a necessary ingredient for the development of deep tropopause folds in two-dimensional contexts, takes place in the present case-study. However, the forcing mechanisms that contribute mostly to subsiding vertical motions over the warm side of the upper-level frontal zone (i.e. with a maximum frontogenetic effect) are, in order of importance, the shear advection and the curvature.

KEYWORDS: Fronts and Atlantic Storm-Track EXperiment, Quasi-geostrophic omega equation, Tropopause fold

1. INTRODUCTION

The structure and the evolution of an exceptionally deep tropopause fold is presented in a set of two companion papers. In Part I (Donnadille et al. 2001) we presented the case-study from the Intensive Observing Period (IOP) 18 (20–24 February 1997) of the Fronts and Atlantic Storm-Track EXperiment (FASTEX, Joly et al. 1999). The initial setting involves a Coherent Tropopause Disturbance (CTD) and an associated Arctic tropopause fold. The confluence episode that results from the phasing up of the tropopause disturbance and a southern ridge ends in the formation of an intense jet streak, the dynamics of which are associated with the development of a polar tropopause fold. A diagnostic analysis has suggested that the final dramatic stratospheric intrusion is the consequence of the vertical superposition of the Arctic and polar tropopause folds. A modelling study has discussed this hypothesis.

In Part I, we stressed the importance of further investigating the mechanisms of upper-level forcing of tropopause folds at the time-scale of the life cycle. This is the purpose of this paper where we propose a diagnostic evaluation of the forcing mechanisms of the vertical motions. Two aspects of the dynamics of upper-level frontogenesis will be considered: (i) the importance of the superposition of confluence and horizontal-shear forcing mechanisms in the presence of cold-air advection, referred to as the Shapiro effect by Rotunno et al. (1994), (ii) modifications of two-dimensional straight-flow conceptual models to cases of three-dimensional curved jet-front systems where the classic dipole model of vertical velocity associated with the curvature effect (Bjerknes and Holmboe 1944) may overlap. In the context of upper-level frontogenesis, the combined roles of confluence and shear are still expected to be elucidated. The transverse

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direct vertical circulation* associated with a pure confluent flow in the entrance of a jet streak can generate the initial stages of a folding, as shown with analytical simulations (Hoskins and Bretherton 1972). However, the pure confluent flow does not allow the stratospheric air to descend below the mid troposphere because the subsidence of the direct vertical circulation is maximized on the cyclonic-shear side of the upper-level jet. Danielsen (1968) and Shapiro (1981) suggested that a combination of confluence and temperature gradient along the jet (which implies cold-air advection and a shearing component and which is also called shear advection) would maximize the subsidence on the warm side of the upper-level front below the jet core and would be necessary for the deepest intrusions to occur. This point was demonstrated by two-dimensional modelling studies based on idealized flow configurations (Keyser and Pecknick 1985a,b): the combined effect of confluence and shear laterally shifts the subsidence branch of the transverse circulation towards the jet axis, resulting in frontogenetical tilting in the prognostic equations for the cross-front potential-temperature gradient and the vertical component of vorticity. In his review of frontal circulations Keyser (1999, pp. 251–254) explains how a positive feedback develops between the vorticity and vertical-motion fields through the horizontal-shear forcing term and how this feedback sets up a transverse and direct ageostrophic circulation that supports tilting in a frontogenetical sense. In this way, the Shapiro effect, considered as a two-dimensional mechanism, helps to account for the observed upper-level frontogenesis, without appealing to three-dimensional arguments involving ageostrophic circulations in the along-front direction. The three-dimensional context of upper-level frontogenesis offers less certainty about the Shapiro effect. First, case-studies involving this effect are lacking in the literature. Second, channel-model simulations of a developing upper-level front in a baroclinic-wave evolution, as in Rotunno et al. (1994) or in Lalaurette et al. (1994), do not develop a large enough phase lag between the geopotential height and thermal patterns to reproduce the Shapiro effect. Using diagnoses of total vertical velocity partitioned into cross- and along-front components, Keyser (1999) suggests the presence of the Shapiro effect in the three-dimensional model simulation of Keyser et al. (1992), but does not reach any conclusions on the positive feedback involving this effect. So, Keyser’s statement (page 254) that ‘The question arises as to whether the Shapiro effect may be found in more realistic models of frontogenesis, and eventually in the real atmosphere’ is still topical.

The structure of the paper is as follows. We use the global operational analyses from the Action des Recherche Petite Échelle et Grand Échelle (ARPEGE) model (described in section 2 of Part I) during the FASTEX period. In section 2 we examine the evolution during IOP 18 of the kinematics aloft, particularly the evolution of vorticity precursors that lead to the formation of a polar jet streak along the upper-level jet. The mechanism that leads to the flow configuration of thermal advection along the jet streak is shown to be associated with the integration of the CTD into a Rossby wave train. This mechanism and its role in the tropopause folding process are discussed in the framework of conceptual models of upper-level frontogenesis, and in terms of the dynamics of jet streaks and of coherent structures on the extratropical tropopause. In section 3, following Jusem and Atlas (1998), we diagnose vertical motions by using a Q-vector partitioning in the natural coordinate system that follows the geostrophic wind. In spite of its quasi-geostrophic framework, the advantage of this partitioning is to

* Throughout the paper, the term vertical circulation shall be used to refer to the vertical-motion field and an associated horizontal field (consistent with mass continuity) in the vicinity of fronts and jet streaks (Keyser 1999). Furthermore, transverse refers to a circulation in a vertical plane oriented normal to the axis of a jet streak or a frontal zone.
evaluate vertical motions associated with the forcing mechanisms previously mentioned: confluence and diffluence, thermal advection by the horizontal geostrophic shear and curvature of the flow. Section 4 discusses the results in the context of present conceptual models and section 5 concludes this set of two companion papers.

2. EVOLUTION OF THE UPPER-LEVEL FLOW CONFIGURATION

From 0000 UTC 21 February to 1200 UTC 22 February, fields of potential temperature on the dynamical tropopause (see Fig. 1 in Part I) show a cold core that is consistently tracked over time from the west of Hudson Bay to the south of Greenland. As this thermal disturbance, the so-called Coherent Tropopause Disturbance (CTD) defined in Part I, comes closer to the polar jet stream, an intense polar jet streak forms along flow south of the CTD. Incorporation of jet streaks in the dynamical framework of the potential-vorticity paradigm (e.g. Hoskins et al. 1985) has recently progressed. Pyle (1997) shows that, depending on their lifetime, CTDs may interact with single or multiple jet streaks. Cunningham and Keyser (1999) show that analytical solutions of barotropic vortex dipoles exhibit characteristic signatures of observed jet streaks. On the basis of a climatological study, Hakim (2000) shows that the tropopause-based disturbances that have nonlinear properties and produce organized vertical motion (nonlinear coherent structures) are, on average, unsteady nonlinear monopolar vortices with locally enhanced gradients of wind speed (jet streaks) embedded in a background flow.

Vorticity precursors of the polar jet streak of the FASTEX IOP 18 case-study are examined. The life cycle is compared with the Shapiro (1982) conceptual model of the evolution of an upper-level jet-front system in a baroclinic wave, and with a revised version proposed by Schultz and Doswell (1999). Diagnoses are presented at 300 hPa to better isolate the signature of the CTD, which on that isobaric surface behaves like a warm anomaly. At 0000 UTC 19 February (Fig. 1(a)) the vorticity features of interest are a ridge at 120°W over the Rocky Mountains and an Arctic trough at (70°N, 100°W). The ridge moves eastwards and the trough moves south-eastwards (Fig. 1(b)). At 0000 UTC 20 February (Fig. 1(c)) the thermal disturbance associated with the CTD (θmax ≥ 314 K) appears in the Arctic trough north-west of the Hudson Bay. By this time the ridge has reached the longitude of the CTD and a synoptic-scale confluent zone sets up. A day later (Fig. 1(d)) the thermal disturbance is over the eastern side of the Hudson Bay, integrating the westerlies and resulting in warm (cold) advection downstream (upstream) of it. At the same time, there is an intensification of the upper-level winds to the south of the CTD, and the formation of a jet streak with wind speeds in excess of 70 m s⁻¹. At 0000 UTC 22 February (Fig. 1(e)) the thermal disturbance has become a warm thermal ridge lying over the cyclonic-shear side of the jet streak and involving cold (warm) advection in the entrance (exit) region of the latter. The flow is roughly straight and has reached its maximum amplitude (wind speeds in excess of 90 m s⁻¹) in the region where the phasing up of the mesoscale trough to the north, and a large-scale ridge to the south of the jet streak occur. At 1200 UTC 22 February (Fig. 1(f)) the thermal disturbance still arranges cold (warm) advection in the entrance (exit) region of the jet streak. The latter has decayed and a midlatitude jet embedded in the large-scale ridge in the southern part of the domain is intensifying.

The first stage in the Shapiro conceptual model involves an upper-level front situated within a confluent north-westerly flow. It is called the equivalent-barotropic stage by Schultz and Doswell (1999) because isentropes are nearly parallel to isohypses. This equivalent-barotropic stage could correspond approximately to Fig. 1(b) in our case-study. During the second stage in the Shapiro conceptual model the thermal trough
Figure 1. Geopotential height (solid lines, every 100 m) and potential temperature (dashed lines, every 4 K) at 300 hPa, winds between 60 and 70 m s$^{-1}$ shaded. (a) 0000 UTC 19 February, (b) 1200 UTC 19 February, (c) 0000 UTC 20 February, (d) 0000 UTC 21 February, (e) 0000 UTC 22 February, (f) 1200 UTC 22 February, location of the vertical cross-section displayed in Fig. 2.
lags behind the trough in the geopotential-height field and involves a maximum of cold advection along the flow at the inflection point in the height field. Such an evolution towards a state of cold advection along the front in north-westerly flow is present in our case-study (Figs. 1(d) and (e)). However, there is no along-front variation in the sign of the thermal advection in the Shapiro conceptual model, a point that is criticized by Schultz and Doswell (1999). In the present case-study the sign of the thermal-advection field reverses at the trough axis of the CTD involving cold (warm) advection upstream (downstream). In the Shapiro conceptual model, geopotential height and thermal waves belong to the same synoptic wave that is close to a quasi-linear Rossby wave (an oscillation due to gradients of potential vorticity, see Hoskins et al. (1985)). In this case-study the thermal disturbance of the CTD is an independent feature that integrates the westerly background flow and superposes its nonlinear effects. It actually corresponds to closed contours of potential temperature at the tropopause with a cold anomaly larger than 10 K (not shown) and a relative-vorticity maximum greater than $10^{-4} \text{ s}^{-1}$ down to 500 hPa (not shown). By non-dimensionalizing the quasi-geostrophic potential-vorticity equation, Hakim (2000) scales an assessment of the nonlinearity as the vorticity ratio of the disturbance and the background flow. Given
the clear distinction between the disturbance and the background flow in our case-
study, nonlinearity can be estimated to be larger than 5, which precludes the linear
wave as a potential model for the disturbance. These differences between the Shapiro
case-study model and the kinematics in the present case-study (i.e. the mesoscale
thermal disturbance integrating the Rossby wave train) call to mind the differences for
growth mechanisms of extratropical cyclones that exist between baroclinic instability
(growth of normal modes, e.g. Simmons and Hoskins (1976, 1977)) and non-modal
baroclinic amplification (Farrell 1984) respectively. As studied by Hakim (2000) the
nonlinearity associated with coherent vortex structures such as the CTD is large,
approaching O(10) for the strongest disturbances.

A vertical cross-section involving the polar jet streak at 1200 UTC 22 February
is presented in Fig. 2 (see location in Figs. 1(f), 4(c) and (f), 5(c) and (f), 6(d)–(f),
7(c) and (f), 8(c) and (f)). The section cuts across its entrance region. The potential-
vorticity field in this cross-section displays a sharp tropopause break along the cyclonic-
shear side of the polar jet core. The surface front is marked by strong ascending
vertical motions and a diabatic-origin maximum of potential vorticity. The potential-
temperature field (not shown) has an upper-level frontal zone that spans the whole
depth of the troposphere from the tropopause break down to the surface front. In
terms of the divergent ageostrophic wind and vertical-velocity components partitioned
in the transverse plane of the cross-section (Ψ-vector technique, Loughe et al. 1995)
the cross-section exhibits a vertical direct circulation with the subsiding branch below
the tropopause break and the ascending branch over the surface front. The prominent
feature of this vertical circulation is its lateral shift towards the anticyclonic-shear side
of the polar jet-streak axis. The subsiding branch, located on the warm side of the
upper-level frontal zone, supports tilting in the frontogenetical sense. The pattern of the
vertical circulation in this cross-section is very similar to the two-dimensional modelled
vertical circulation in the simulation of Keyser and Pecknick (1985a,b) devoted to
the investigation of the Shapiro effect. However, this result is not a demonstration of
the existence of the Shapiro effect because the vertical circulation displayed in Fig. 2
includes the contributions of the confluence and of the shear-advection terms.

3. QUASI-GEOSTROPHIC OMEGA DIAGNOSES

Following Jusem and Atlas (1998) we propose a partitioning of the Q-vector
into along- and cross-isohypse components on an isobaric surface. In addition to
the evaluation of vertical-motion forcing mechanisms as confluence/diffusence and
curvature, the interest in such a partition is to quantitatively assess the vertical-motion
forcing due to the thermal advection by the horizontal geostrophic shear. The reader is
referred to Jusem and Atlas (1998) for a physical basis and a complete derivation of the
quasi-geostrophic equation, with pressure coordinate, P, for the vertical velocity, \( \omega \), in
terms of the natural coordinate system of the isohypsises, \( \phi \). Only a summary is made in
this paper. The \( \omega \)-equation is written as:

\[
\sigma \nabla^2 \omega + f_0^2 \frac{\partial^2 \omega}{\partial P^2} = F = -2 \nabla \cdot Q
\]

where \( F \), the geostrophic forcing, is equal to twice the convergence of the Q-vector,
\( \sigma = -\alpha_0 \Theta^{-1} \frac{\partial \Theta}{\partial P} \) is a stability parameter, \( \alpha_0 \) is a reference-state specific volume
and \( \Theta \) is a reference-state potential temperature, both depending on pressure only. The
dependence of \( Q \) on the fields of geopotential height, \( \phi \), and specific volume, \( \alpha \), takes the form
\( Q = -\nabla \cdot \nabla \alpha \) and involves the kinematic constraint of non-divergence on
the geostrophic wind \( \mathbf{V}_g \), namely, \( \nabla \cdot \mathbf{V}_g = 0 \). The natural coordinate system of the isohypses is defined by the right-handed triplet of unit vectors \((\mathbf{t}, \mathbf{n}, \mathbf{k})\) where \( \mathbf{t} \) is collinear to the local axis, \( s \), parallel to the geostrophic wind, \( \mathbf{n} \) is collinear to the local axis, \( n \), normal to the geostrophic wind, and \( \mathbf{k} \) is the vertical unit vector.

If \( K_s \) is the curvature of the geopotential contour lines, \( K_n \) is the curvature of the lines orthogonal to the geopotential contour lines, and \( s_g \) is the geostrophic wind speed, the \( Q \)-vector in the natural coordinate system can be written as the sum of four components (Jusem and Atlas 1998):

\[
Q_{\text{crst}} = -n s_g K_n \frac{\partial \alpha}{\partial n} = n \frac{K_n}{f_0} \frac{\partial \phi}{\partial n} \frac{\partial \alpha}{\partial n}
\]

\[
Q_{\text{curv}} = -ts_g K_s \frac{\partial \alpha}{\partial n} = t \frac{K_s}{f_0} \frac{\partial \phi}{\partial n} \frac{\partial \alpha}{\partial n}
\]

\[
Q_{\text{shev}} = -n s_g \frac{\partial \alpha}{\partial s} = n \frac{1}{f_0} \frac{\partial^2 \phi}{\partial n^2} \frac{\partial \alpha}{\partial s}
\]

\[
Q_{\text{alst}} = -t s_g \frac{\partial \alpha}{\partial s} = t K_n s_g \frac{\partial \alpha}{\partial s}
\]

\( Q_{\text{crst}} \) is the cross-stream component that represents the effect of confluence and diffuence of the geostrophic wind, it is the physical basis of the four-quadrant conceptual model of vertical velocity around a straight jet streak (Bjerknes 1951; Riehl et al. 1952). \( Q_{\text{curv}} \) is the along-stream component that represents the curvature effect according to which a downstream increase (decrease) in the cyclonic curvature of the isohypses induces subsidence (ascent). The two other additional \( Q \)-vector components come in a baroclinic atmosphere when there is a significant intersection of isotherms and isohypses on an isobaric surface. \( Q_{\text{shev}} \) is the cross-stream component that represents the thermal advection by the horizontal geostrophic shear; it is called the shear advection component of \( Q \). Its effect, when combined with that of the confluence, represents the so-called Shapiro effect. \( Q_{\text{alst}} \) represents the along-flow stretching/contraction of isotherm spacing. Jusem and Atlas (1998) show that this process, which always operates in the presence of confluence/diffuence, is negligible in their case-study. It is also verified in our case-study so that no map of \( Q_{\text{alst}} \) and its associated vertical velocity will be shown. Each of the four quasi-geostrophic forcing terms can be treated independently as solutions to the left-hand side of the \( \omega \) equation, with the sum of the solutions for vertical motions being equal to \( \omega_{\text{tot}} = \omega_{\text{crst}} + \omega_{\text{curv}} + \omega_{\text{shev}} + \omega_{\text{alst}} \).

\( (a) \) Total forcing

The degree of realism of the geostrophic forcing of vertical motion is assessed by a comparison with the vertical motion derived from a kinematic method and from a more accurate nonlinear balance than the quasi-geostrophic balance. For the kinematic method, vertical motions are computed by integrating the equation of continuity in pressure coordinates and adjusting by a linear correction of the divergence (O’Brien 1970). For the more accurate nonlinear balance we use the Davies-Jones (1991) alternative balance (AB) omega equation. In this equation horizontal and vertical derivatives of the wind are approximated by those of the non-divergent and geostrophic winds, respectively. The dynamical forcing is independent of the divergent part of the wind, i.e.

* The forcing associated with \( Q_{\text{alst}} \) (see Eq. (2.17) in Jusem and Atlas (1998)) is the sum of two terms involving \((\partial \alpha/\partial s)\) and \((\partial^2 \alpha/\partial s^2)\), both of which we have checked to be small, and which, furthermore, cannot be maximum at the same location.
Figure 3. Vertical velocity at 500 hPa at 0000 UTC 22 February: (a) from the kinematic method (every 1 \( \mu \text{b s}^{-1} \), shaded for subsiding motions, the thick isoline is the 70 m s\(^{-1}\) isotach at 300 hPa), (b) as for (a) but for the dynamical contribution derived from the alternative-balance approximation, (c) as for (a) but for the dynamical contribution derived from the quasi-geostrophic approximation.

the vertical circulation. Furthermore, the advantage of the AB assumption is that the material derivative of the thermal wind imbalance is zero, which is less restrictive than setting to zero the thermal imbalance itself (see Davies-Jones 1991, page 507). The complete expression of the AB omega equation used here is also given by Mallet et al. (1999). The computations are made on gridded data on a polar stereographic projection, using a 75 \( \times \) 75 km horizontal resolution for both the kinematic method and the AB omega equation, and a 200 \( \times \) 200 km horizontal resolution for the terms of the quasi-geostrophic equation. Implicitly included in the kinematic method, the diabatic forcing is ignored in the AB and quasi-geostrophic (QG) omega equations; this point is not considered as a drawback here as we are mostly interested in the patterns of subsiding vertical motions. Figure 3 presents a comparison of the three fields of vertical motions in the mid troposphere at the time when the intense polar jet streak approaches the base of the geopotential-height trough (see Fig. 1(e)). Vertical motions computed kinematically are the most intense and contain mesoscale information. Part of this information is lost in the field of the AB vertical motion, owing to the absence of the diabatic forcing contribution and to the AB approximation. However, the AB \( \omega \) field resembles the kinematic one as far as the depiction of large structures is concerned. The field of quasi-geostrophic vertical motion shows an additional smoothing, but again the largest structures seen with the kinematic method are represented. Other comparisons of the QG and kinematic omega fields during the life cycle are analysed in section 4.
Figure 4. Left-hand panels: Q-vector component at 500 hPa derived from the confluence/diffuence term $Q_{\text{con}}$, geopotential lines (solid, every 100 m), the thick isoline is the 40 m s$^{-1}$ geostrophic isolach. The strongest Q-vector shown is $3 \times 10^{-12}$ m$^2$kg$^{-1}$s$^{-1}$. Right-hand panels: quasi-geostrophic omega component at 500 hPa derived from the confluence/diffuence term, $\omega_{\text{con}}$ (every 0.5 µb s$^{-1}$, shaded for subsiding motions), the thick isoline is the 300 hPa 70 m s$^{-1}$ observed isolach, geopotential lines (solid, every 150 m). (a), (d) 0000 UTC 21 February. (b), (e) 0000 UTC 22 February. (c), (f) 1200 UTC 22 February.

(b) The four-quadrant model

In Fig. 5 are portrayed the 500 hPa fields of geopotential height and specific volume (proportional to the temperature on an isobaric surface). The strengthening of the gradient of the specific volume in the region of the polar jet streak characterizes the intense ongoing mid-tropospheric frontogenesis. Figures 4(a)–(c) are for $Q_{\text{con}}$ associated with confluence/diffuence; Figures 5(a)–(c) are for $Q_{\text{shdv}}$ associated with
shear advection; Figures 7(a)–(c) are for \( Q_{\text{curv}} \) associated with curvature; Figures 4(d)–(f), 5(d)–(f), 7(d)–(f) show the corresponding vertical velocities. In this way, patterns of the flow involving either confluence/diffuence, thermal advection by the geostrophic shear, or curvature can be qualitatively assessed.

\( Q_{\text{crst}} \) is significant in regions of confluence and diffuence. It takes the direction of \( \nabla \phi \) (\( -\nabla \phi \)) in confluent (diffuent) regions. At 0000 UTC 21 February \( Q_{\text{crst}} \) is intensifying in the region of the incipient polar jet streak to the west of the Great Lakes. At 0000 UTC 22 February this component is large in the entrance region of the polar jet streak and is intensifying in its exit region east of Newfoundland. At 1200 UTC 22 February both the entrance and exit regions of the polar jet streak have large \( Q_{\text{crst}} \) vectors. The contribution to the vertical motion \( \omega_{\text{crst}} \) that comes from the partial forcing \( -2\nabla \cdot Q_{\text{crst}} \) (Figs. 4(d)–(f)) displays a four-quadrant pattern along the polar jet streak and is similar to the ‘four-quadrant model’ of a straight jet streak (Bjerknes 1951; Riehl et al. 1952): it appears from 0000 UTC 22 February and is most strongly defined at 1200 UTC 22 February. In a conceptual way it represents the vertical transverse direct (indirect) ageostrophic circulation associated with the convergence (divergence) pattern in the entrance (exit) region of a straight jet streak (Uccellini and Johnson 1979). An alternative conceptual model is the pattern of vertical motion associated with the advection of a cyclonic–anticyclonic dipole of isobaric relative vorticity straddling the jet-streak axis. As noted by Cunningham and Keyser (1999) departures from such conceptual models come from certain observed features of jet streaks, such as the anisotropy of the wind field in the along-stream direction and the asymmetry of the relative-vorticity field in the cross-stream direction. The present polar jet streak also exhibits such complicating features. For example, the cyclonic vortex is typically stronger and more compact than the anticyclonic vortex (not shown), which is reflected in the stronger dipole of \( \omega_{\text{crst}} \) along the cyclonic-shear side of the jet streak (Fig. 4(f)). Modifications of this popular four-quadrant model that account for along-jet thermal advection and flow curvature are now investigated by considering the effects of the other contributions to \( Q \).

(c) The Shapiro effect

The effect of thermal advection by the horizontal geostrophic shear is illustrated in Fig. 5, representing the cross-stream \( Q_{\text{shdy}} \) vectors and the corresponding vertical velocity, \( \omega_{\text{shdy}} \). At 0000 UTC 22 February, the \( Q_{\text{shdy}} \) vectors are significant in two regions along the cyclonic-shear side of the polar jet streak. These two regions are separated by the axis of the 500 hPa geopotential-height trough whose progression is linked with that of the CTD (Fig. 1 in Part I). At 0000 UTC 22 February (Fig. 5(b)) the cold advection is large on the cyclonic-shear side of the polar jet streak upstream of the CTD and diminishes northward. It gives rise to a meridional dipole of \( \omega_{\text{shdy}} \) (Fig. 5(e)) with a prominent subsiding pole centred on the polar jet axis and a much weaker pole of ascending motion to the north ((58°N, 61°W), scarcely visible). The warm advection intensifies on the cyclonic-shear side of the polar jet streak downstream of the CTD, its contribution to \( \omega_{\text{shdy}} \) appearing as a second dipole with ascent along the jet at the exit region (48°N, 50°W), and with descent on the cyclonic-shear side of the polar jet streak (56°N, 55°W). At 1200 UTC 22 February (Figs. 5(c) and (f)) the asymmetry in the temperature advection pattern is smaller and a four-quadrant distribution of \( \omega_{\text{shdy}} \) appears at the cyclonic-shear flank of the polar jet streak. The south-west to north-east dipole of \( \omega_{\text{shdy}} \) lying in the entrance region depicts a transverse indirect circulation. There is a partial overlap of the subsiding branch of the latter with the subsiding branch of the direct circulation associated with \( \omega_{\text{crst}} \), which reinforces the subsidence below the
jet. Thus, this Q-vector partition makes it possible to point to the two-cell transverse circulation associated with confluence plus shear that is necessary to create a deep fold (Danielsen 1968; Shapiro 1981).

(d) The curvature effect

Curvature and shear potential-vorticity components, analysed with a shear-curvature partitioning of the potential-vorticity field in natural (streamline following) coordinates (Bell and Keyser 1993), are displayed on the 300 K surface for the start and the end of the life cycle (Fig. 6). It shows that, beginning with a predominant curvature component,
Figure 6. Shear-curvature partitioning of the potential-vorticity field at $\theta = 300$ K in natural (streamline following) coordinates (Bell and Keyser 1993). (a) and (d) total potential vorticity at 0000 UTC 20 February and at 1200 UTC 22 February, (b) and (e) curvature component, (c) and (f) shear component.
there is a tendency in the internal conversion in the potential-vorticity field to increase (decrease) the shear (curvature) component from the start to the end of the life cycle. Note in this connection, how the maximum of the shear component of the isentropic potential-vorticity field at 1200 UTC 22 February (Fig. 6(f)) is just within the dipole of $\omega_{\text{crist}}$ on the cyclonic-shear side of the polar jet streak. The increase in the shear component at the expense of the curvature component leads to the formation of closed isolines of the shear component of PV on the cyclonic-shear side of the jet streak and brings the system closer to the conceptual model of a dipole of vorticity straddling a straight jet streak (Cunningham and Keyser 1999).
The curvature component, $Q_{\text{curv}}$, and its associated vertical velocity, $\omega_{\text{curv}}$, are displayed in Fig. 7. $Q_{\text{curv}}$ is directed downstream (upstream) of the geostrophic wind in cyclonic (anticyclonic) regions. The curvature effect is proportional to the wind speed, $V$, and to the curvature of the geopotential contour lines, $K_s$; as seen with the analytic expression of the curvature forcing $(-2V \cdot \mathbf{Q}_{\text{curv}}$, Eq. (2.18) in Jusem and Atlas (1998)) or with a simple conceptual model of vorticity advection in natural coordinates (with the vorticity expression $\eta = -(\partial V/\partial n) + V K_s$). In this way, a slight curvature in a jet streak may be sufficient to have an effect on vertical velocity as large as the confluence and shear forcings. This is indeed the case here, as a well defined dipole of vertical velocity follows the movement of the CTD and polar jet streak and intensifies as they advance (Fig. 7(d)–(f)). In itself the dipole model of the curvature effect is a classic result derived from the paper by Bjerknes and Holmboe (1944) and updated, for instance, by Uccellini (1990). The interest in this case-study is the interpretation in a three-dimensional context of the combined forcings by confluence, shear and curvature.

4. DISCUSSION

Figure 8 displays the fields of the QG dynamic vertical motion, $\omega_{\text{tot}}$, and the kinematic vertical motion for the time period considered. Except in ascending regions where latent heat can enhance the vertical velocity, the QG dynamic vertical motion compares reasonably well with the kinematic vertical motion. In subsiding regions, given the separate contributions of $\omega_{\text{crst}}$, $\omega_{\text{shdv}}$ and $\omega_{\text{curv}}$, it is not surprising that the curvature effect dominates the QG vertical-motion field. It is worth noting that the vertical motion induced by the curvature effect does not cancel any feature of the vertical motion induced by confluence and shear advection, but on the contrary adds a large contribution to these fields. We will give a commentary on this point at the end of this section regarding the triggering of the cyclogenesis.

As stated previously, the key point for the efficiency of upper-level frontogenesis is to maximize the subsidence beneath the jet axis on the warm-air side of a developing upper-level frontal zone. In the two-dimensional framework, a candidate mechanism is the Shapiro effect. Looking at Figs. 4 and 5 it is clear that the FASTEX IOP 18 case-study offers an example of the Shapiro effect taking place in the real atmosphere. The subsiding pole of $\omega_{\text{shdv}}$ in the entrance region is always south of the one of $\omega_{\text{crst}}$ and shifts the transverse circulation towards the jet-streak axis. Now, it is important to come back at this stage to the life-cycle aspect of the tropopause-fold development and to the feature that initiates and configures the upper-level flow during FASTEX IOP 18, that is to say, the CTD. Once this feature, that was embedded in the stratospheric reservoir, integrates the southern Rossby wave train, all forcing fields simultaneously exist, i.e. the confluence field that brings it closer to the incipient jet stream, the thermal advection field that involves cold (warm) advection upstream (downstream) of it, and the curvature field. So, there is clear evidence in this case that the forcing terms of confluence, shear advection and curvature, although separable for diagnostic purposes, have to be considered as originating from the same feature, the CTD, in order to investigate the observed upper-level frontogenesis.

Three-dimensional channel-model simulations of baroclinic waves (Lalaurette et al. 1994; Rotunno et al. 1994) do not develop the Shapiro effect in the early stages of the wave evolution: in such simulations the initial and analytical state cannot include the nonlinearity brought by features such as CTDs. It remains for future research to investigate whether FASTEX IOP 18-like cases are more numerous than has been thought. Finally, it is important to note in Figs. 4 and 8 the asymmetry that exists
between the entrance and the exit regions of the polar jet streak in terms of the forcing mechanisms of the vertical motion. Diffuence, horizontal shear in the presence of warm advection and positive advection of curvature all add contributions to force ascending motions on the cyclonic-shear side of the exit region of the polar jet streak. Although such an investigation is out of the scope of the present paper, it is plausible that diagnoses developed here could help in the understanding of how explosive surface cyclogenesis is triggered during FASTEX IOP 18.
5. Conclusions

This study, part of a set of two companion papers, has documented the life cycle of a very deep tropopause fold (820 hPa) during IOP 18 (20–24 February 1997) of FASTEX. Part I described the initial setting which involves a CTD and an associated Arctic tropopause fold. The confluence episode that results from the phasing up of the CTD and a southern ridge ends in the formation of an intense jet streak, the dynamics of which are associated with the development of a polar tropopause fold. A diagnostic analysis has suggested that the final dramatic stratospheric intrusion is the consequence of the vertical superposition of the Arctic and polar tropopause folds. A modelling study has discussed this hypothesis.

In Part II, we have investigated the dynamics of the upper-level flow configuration to assess the contributions of the vertical-velocity forcing terms. The goal was to revisit two-dimensional straight-flow conceptual models of upper-level frontogenesis and their modifications for cases of three-dimensional curved jet-front systems. Based on the Jusem and Atlas (1998) Q-vector partition in the natural coordinate system that follows the geostrophic wind, a diagnostic study investigated the forcing terms of the vertical motion by confluence, horizontal shear in the presence of thermal advection, and curvature. Results indicated that the curvature effect does not cancel any feature of the two-dimensional forcing terms by confluence and shear but, on the contrary, adds a large contribution to these fields.

Further, results showed that a Shapiro effect actually takes place (combination of confluence and horizontal shear in the presence of cold advection) in the case-study. This effect is considered in the literature as a necessary ingredient for the occurrence of the deepest tropopause folds. It has been successfully modelled with two-dimensional simulations (e.g. Keyser and Pecknick 1985a,b) but to our knowledge not with three-dimensional ones (e.g. Lalaurette et al. 1994; Rotunno et al. 1994). In the case-study, this effect was shown to be dependent on the existence and the spatial scale of the CTD and, therefore, must be considered as the result of a three-dimensional mechanism of the observed upper-level frontogenesis. The failure of three-dimensional channel-like simulations to resolve this process is probably a consequence of the weak degree of nonlinearity involved in the initial atmospheric state normally used for simulations of normal-mode-growing cyclogenesis. During FASTEX IOP 18 the degree of nonlinearity involved in the integration of the CTD into a Rossby wave train is higher. A question that also arises is how representative is the FASTEX IOP 18 upper-level flow configuration of the life cycle of midlatitude baroclinic waves.

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