Effects of a low-level precursor and frontal stability on cyclogenesis during FASTEX IOP17

By ISABELLE MALLET¹, PHILIPPE ARBOGAST¹, CHRISTOPHE BAEHR¹, JEAN-PIERRE CAMMAS² and PATRICK MASCART²*

¹Météo-France, Centre National de Recherches Météorologiques, France
²Laboratoire d’Aérologie, France

(Received 26 August 1998; revised 14 April 1999)

SUMMARY

The Fronts and Atlantic Storm-Track EXperiment (FASTEX) has provided comprehensive data to document the life-cycle of secondary frontal cyclones over the North Atlantic, and to improve the understanding of cyclogenesis mechanisms. This study analyses the processes leading to the triggering of a particularly well-sampled frontal cyclone that developed on the trailing cold front of a mature primary cyclone during FASTEX Intensive Observation Period 17 (IOP17) between 16 and 20 February 1997.

The case features both a classical low-level vorticity strip along the primary front where the frontal cyclone develops and, more unexpectedly, a pre-existing continental surface low further west, the importance of which is revealed in the pilot study of Arbo gast and Joly. In order to study possible cyclogenetic contributions along the primary front, the domain-independent vorticity-divergence attribution technique of Bishop is used to partition the flow into contributions from the continental low and from the large-scale environment. Results indicate that the frontal cyclone is triggered where and when along-front stretching decreases below the theoretical critical threshold of Bishop and Thorpe. It is shown that the cyclone is developing under normalized strain less than 1, in agreement with other case-studies. Finally, the role of the continental low is investigated using a series of numerical forecasts made from different initial conditions manipulated with a potential-vorticity inversion tool. The wind attributable to the continental low accounts for roughly 30% of the along-front stretching decrease. The continental low also has a frontolytic effect on the western part of the primary front where its thermal advection pattern favours cyclogenesis.

In general terms, the frontal-cyclone triggering seems to support the Bishop and Thorpe theoretical framework of frontal instability in low-strain areas. However, a more complex picture arises when considering the role of the continental surface low, which also acts as a low-level precursor.

KEYWORDS: Deformation FASTEX Frontogenesis Potential-vorticity inversion Secondary cyclone Stretching

1. INTRODUCTION

North Atlantic mid-latitude secondary cyclones are waves which develop on the trailing cold front of a mature large-scale baroclinic wave within the oceanic storm track. These frontal cyclones are sub-synoptic systems with typical scales of the order of 1000 km (Renfrew 1995). This study analyses the processes leading to the triggering and development of a frontal cyclone observed during the Fronts and Atlantic Storm-Track EXperiment (FASTEX) field phase (Joly et al. 1999). The case chosen corresponds to FASTEX Intensive Observation Period 17 (IOP17); it is a particularly well-sampled event that FASTEX participants refer to as ‘The FASTEX cyclone’ (16–20 February 1997). The storm was triggered during the 6 h period from 0600 UTC to 1200 UTC 17 February, and eventually developed into a 941 hPa surface-low centre by 0000 UTC 20 February. The synoptic and dynamic overview of the first two days of the life cycle is presented in Cammas et al. (1999).

The very first examination of this event by Arbo gast and Joly (1998b) revealed that the triggering of the IOP17 secondary cyclone is not linked in a straightforward way to an upper-tropospheric disturbance but, unexpectedly, is related to a low-level precursor. Using a potential-vorticity-inversion (PV-inversion) tool, they identified this low-level precursor as a pre-existing surface low, initially located over the continental USA in the Great Lakes area. The present paper discusses further the triggering mechanisms.

* Corresponding author: Laboratoire d’Aérologie, UMR CNRS/UPS 5560, Observatoire Midi-Pyrénées, 14 Av. Edouard Belin, 31400 Toulouse, France. e-mail: masp@aero.obs-mip.fr
of the FASTEX cyclone using the domain-independent vorticity-divergence attribution method developed by Bishop (1996a,b). We try to identify the remote effects linked to the aforementioned continental surface low, and examine how the wind field contribution related to this low decreases the stability of the primary cold front where the frontal cyclone is eventually triggered.

Theoretical studies by Bishop (1993a,b) and Bishop and Thorpe (1994a,b) have pointed out the importance of the large-scale horizontal deformation for frontal wave development. Indeed, along-front stretching deformation can suppress the growth of frontal waves, if strong enough. The method developed by Bishop (1996a,b) allows frontal stability signatures to be estimated in real cases of frontal waves, by providing a way of partitioning the observed wind field into environmental and frontal components. Thus the environmental along-front stretching in the frontal area can be compared to its theoretical counterpart, the large-scale deformation of the basic-state field. The Bishop domain-independent attribution tool has been applied by Renfrew et al. (1997) and Rivals et al. (1998) to several cases of frontal waves observed over the Atlantic. They found that the range of environmental along-front stretching was consistent with the theoretical critical-threshold value of Bishop and Thorpe (1994b). Very recently, Chaboureau and Thorpe (1999) have proposed a more general approach which extends the wind partitioning of Bishop (1996a) to the temperature field by assuming a balance condition; and they used this technique to survey the impact of along-front temperature gradient and wind shear on cyclogenesis for a selection of FASTEX cases.

(a) Overview of IOP17 life cycle

The data used for the present case-study were provided by the optimal interpolation analysis scheme of the Météo-France ARPEGE* operational model (Courtier et al. 1991). The ARPEGE spectral model is based on a stretched geometry (Schmidt 1977; Courtier and Geleyn 1988). During FASTEX, it used a T149 spectral truncation on the stretched sphere, a stretching factor $C$ of 3.5 that provides a resolution of about 30 km over France and 80 km near Newfoundland, and 27 eta ($\eta$) levels in the vertical.

During the 4-day period from 1200 UTC 16 February to 0000 UTC 20 February 1997, the IOP17 frontal cyclone evolution broadly comprises five distinct phases: (i) an incipient phase ending at about 0000 UTC on 17 February; (ii) a triggering phase ending at about 1200 UTC on 17 February; (iii) a first development phase ending at about 1200 UTC on 18 February; (iv) a second development phase ending at about 1200 UTC on 19 February; and finally (v) a mature phase. The scope of this paper is restricted to the incipient and triggering phases, (i) and (ii). Only a summary of the relevant points of the overview of Cammas et al. (1999) is given in this section.

Early in the incipient phase at 1200 UTC 16 February (Fig. 1(a)) the main baroclinic zone is the cold front trailing from south-east of Greenland to Florida, the parent low being centred south of Iceland. This baroclinic zone is associated with an intense upper-tropospheric jet stream (shown at 1800 UTC 16 February in Fig. 4(a)). Two other important features are noticed further west. A continental surface low is located south of the Great Lakes. Aloft, an upper-level trough over the central USA contains a tropopause height anomaly (see the PV maximum on the 300 K isentropic surface in Fig. 1(a)) which seems favourably located to interact positively with the continental surface low. During the incipient phase, after 0000 UTC 17 February, the upper-level anomaly moves towards the cyclonic side of the upper-level jet, increases the jet intensity (Fig. 4(b)), and triggers a transverse and direct ageostrophic circulation (TDA) in the jet entrance region.

* Action de Recherche Petite Echelle Grande Echelle.
Figure 1. IOP17 incipient phase synoptic situation (operational analysis) showing: sea-level pressure (bold lines; 5 hPa intervals); 850 hPa wet-bulb potential temperature (grey lines; 1 K intervals from 281 to 286 K); potential vorticity on 300 K isentropic surface (bold dashed lines above 1 PVU, shaded from 1 to 2 PVU; $1 \text{ PVU} = 10^{-6} \text{K m}^{-2}\text{s}^{-1}\text{kg}^{-1}$): (a) at 1200 UTC 16 February (initial conditions for experiments A, B and C), the heavy line indicates the position of vertical cross-sections in Fig. 2; (b) at 1200 UTC 17 February.

(Cammas et al. 1999). By 1200 UTC 17 February (Fig. 1(b)), the continental surface low has moved offshore over the Atlantic, and has merged into a broad low-pressure area out of which the IOP17 surface low has emerged (1018 hPa at 36°N, 61°W).

The roles of both the upper-level anomaly and the continental surface low in triggering the cyclogenesis are investigated by Arbogast and Joly (1998b). A variational PV-inversion tool (briefly explained in appendix A) is used to modify initial conditions of the ARPEGE spectral model, and perform three experiments described in Table 1. As explained in appendix A, these experiments use a T63 spectral truncation (coarser than the T149 truncation of the operational ARPEGE model) that provides a horizontal
TABLE 1. DATASETS USED IN EXPERIMENTS

<table>
<thead>
<tr>
<th>Exp. O</th>
<th>Exp. A</th>
<th>Exp. B</th>
<th>Exp. C</th>
</tr>
</thead>
<tbody>
<tr>
<td>6 hourly operational analyses (T149L27)</td>
<td>Reference forecast (T63L19) from operational analysis</td>
<td>Forecast (T63L19) initialized without upper-level anomaly</td>
<td>Forecast (T63L19) initialized without continental low</td>
</tr>
<tr>
<td>60 h range, from 1200 UTC 16 Feb.</td>
<td>60 h range, from 1200 UTC 16 Feb.</td>
<td>60 h range, from 1200 UTC 16 Feb.</td>
<td></td>
</tr>
</tbody>
</table>

resolution close to 200 km, and 19 vertical levels. The main results from these three experiments, run from several different initial states (Table 1 and Fig. 2), are summarized here. Experiment A is the reference simulation, it is initialized with the ARPEGE operational analysis and correctly captures IOP17 cyclogenesis triggering and development (Fig. 2). Experiment B investigates the sensitivity of IOP17 cyclogenesis to the upper-level anomaly. Initial conditions are modified using the inversion tool to withdraw the upper-level PV anomaly. In this case a weak surface-pressure wave shows up, but cyclogenesis is not really triggered (Fig. 2). Arbogast and Joly (1998b) show that the cyclogenesis mechanism is similar to a Type B development (Petterssen and Smebye 1971), but only operates after some delay, increasing the deepening rate when the IOP17 triggering phase is completed. They suggest that the upper-level anomaly first contributes to the maintenance of the continental surface low during its movement from the Great Lakes to the eastern Atlantic coast, and only later interacts baroclinically with the incipient IOP17 low. Accordingly, the direct precursor seems to be the moving continental surface low, not the upper-level anomaly. This conclusion is supported by experiment C, in which the continental surface low is discarded from the initial conditions but the upper-level anomaly is kept in place. In this case, some cyclogenesis occurs but the deepening remains weak (Fig. 2). Thus, both the upper-level anomaly and the continental surface low are required components in the IOP17 triggering.

As the baroclinic interaction between upper and lower features occurs after the triggering of the IOP17 cyclone, it allows us further insight into the mechanisms linking the low-level precursor to the cyclogenesis using the Bishop technique (1996a,b).

(b) Methodology

The domain-independent vorticity and divergence attribution technique of Bishop (1996a,b) allows an important issue of the frontal stability theory to be addressed, by separating the action of the environment on the front from the local action of the front on itself. In the present work, it is also used to isolate the remote actions of the continental surface low in the front area. The technique allows the wind field to be partitioned into one part due to any local structure and one part due to the remaining environmental flow. Diagnoses of wind derivatives can then be performed on each part of the flow. A frontal flow and an ambient flow are computed by defining a ‘frontal box’ enclosing the vorticity and divergence signatures of the frontal strip. Since we follow the same approach as in Rivals et al. (1998), only a short summary is given here (see appendix B). For all the attribution calculations, the analysed or forecast fields are interpolated onto a polar stereographic conformal grid (75 km horizontal resolution, 19 pressure levels ranging from 1000 to 100 hPa).

The attribution technique is applied to 950 hPa winds, on which the signatures of the continental surface low and of the surface cold front (geopotential, vorticity and divergence) can be seen until 1200 UTC 17 February on ARPEGE operational analyses. Renfrew et al. (1997) and Rivals et al. (1998) have shown that the technique is not
very sensitive to the level used in spite of possible boundary layer effects. Here we have checked that our results are qualitatively similar when 900 hPa winds are used. In Fig. 3, the trailing primary cold front of the parent low is depicted by a strip of cyclonic vorticity extending across the Atlantic from south-east of Greenland to Florida. The western tip of this strip lies under the anticyclonic side of an upper-level jet entrance (Fig. 4) where the TDA circulation becomes stronger (notice the change in 300 hPa along-stream gradient of wind speed between Fig. 4(a) and 4(b)). The vorticity (Fig. 3) and divergence (not shown) structure associated with the continental surface low, moving along the 40°N parallel, is the sole coherent feature in the vicinity of the primary cold front. Elsewhere in the surrounding areas, vorticity and divergence patterns are weak, of small scale and random; they show no time continuity between successive 6 h analysis slots. Although
950 hPa relative vorticity (contour interval 2.5 × 10⁻⁵ s⁻¹; shaded between 10 and 15 × 10⁻⁵ s⁻¹, dashed for negative values) overlaid with wind vectors (scale at bottom right), with positions of the 3 boxes BW, BE and B₁ (in solid, dash-dotted and dashed lines respectively): (a) at 1200 UTC 16 February; (b) at 1800 UTC 16 February; (c) at 0000 UTC 17 February; (d) at 0600 UTC 17 February, with IOP17 low centre marked by L. See text for further details.

attraction is a two-dimensional technique on a pressure level, the 950 hPa winds include effects from other levels; for instance the low-level cyclonic circulation associated with upper-level PV anomalies (Hoskins et al. 1985), or the vertical ageostrophic circulations in the entrance or exit regions of upper-level jet streaks (Keyser and Shapiro 1986).

Our objectives in the forthcoming sections are, on one hand, to assess the stability of the primary front under the influence of its straining environment and, on the other hand, to discuss the role of the continental surface low and clarify the mechanisms by which this low favours IOP17 cyclogenesis off the east coast of North America. We will proceed in three steps:

(i) In the first step, the strain along the primary front is analysed in two areas following the approach by Rivals et al. (1998). A rectangular box BW is defined (Table 2 and Fig. 3) to encompass the western part of the primary surface front where the IOP17
cyclone develops. A control box $B_E$, shaped as a segment of an annulus, is located over the slightly bent eastern part of the primary front trailing directly south of the North Atlantic parent low, where no cyclogenesis occurs. The environmental wind is defined by successively considering the winds associated with the vorticity and divergence elements out of box $B_W$ or $B_E$. For this analysis, the ARPEGE operational analyses fields (Table 1) are used.

(ii) The second-step objective is to examine the role of the continental surface low. A box containing the continental low ($B_l$, Table 2) is used to make two tests excluding wind contributions from the continental surface low. First, an environmental wind field is computed from the ARPEGE operational analyses by masking vorticity and divergence elements from both box $B_l$ (continental surface low) and $B_W$ (where IOP17 cyclogenesis is triggered), and the results are compared to those of step (i). However, this comparison only documents the instantaneous remote action of the continental surface low on the low-level baroclinic area associated to the primary front. Therefore, a new strain analysis is performed using forecast wind fields from experiment C (Table 1), and computing the environmental wind by excluding box $B_W$ (western part of the front). Comparing the results to the previous ones allows us to discuss the time-integrated evolution of the primary front stability when the continental surface low is excluded. The same test is also performed using forecast wind fields from experiment B (Table 1) to confirm that the upper-level anomaly helps to maintain the continental surface low.

(iii) The final step tries to clarify the remote-action mechanisms linking the continental surface low to the primary front baroclinic zone. For this step, the wind field contribution attributable to the continental surface low (box $B_l$) is used to estimate its remote action in terms of stretching, frontogenesis, and thermal advection in the area of the front.

The geometric parameters of boxes $B_W$, $B_E$ and $B_l$ (size and shape) are summarized in Table 2 and their locations are shown on Fig. 3. The long sides of boxes $B_W$ and $B_E$ are aligned with vorticity contours (Fig. 3). Boxes $B_W$ and $B_E$ are of equal area to allow
TABLE 2. CHARACTERISTICS OF BOXES \( B_W \), \( B_E \) AND \( B_I \)

<table>
<thead>
<tr>
<th>Box</th>
<th>Isolated domain</th>
<th>Exp.</th>
<th>Date and time (UTC)</th>
<th>Dimension (km) width</th>
<th>Dimension (km) length</th>
<th>Type</th>
</tr>
</thead>
<tbody>
<tr>
<td>( B_W )</td>
<td>western part of the surface front</td>
<td>O</td>
<td>1200 to 1200 Jan.</td>
<td>750</td>
<td>2250</td>
<td>rectangle</td>
</tr>
<tr>
<td></td>
<td></td>
<td>A</td>
<td>1200 to 1200 Jan.</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>B</td>
<td>1200 to 1200 Jan.</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>C</td>
<td>1200 to 1200 Jan.</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>( B_E )</td>
<td>eastern part of the surface front</td>
<td>O</td>
<td>1200 to 1200 Jan.</td>
<td>750</td>
<td>2250</td>
<td>annulus sector</td>
</tr>
<tr>
<td>( B_I )</td>
<td>continental low</td>
<td>O</td>
<td>1200 to 1800 Jan.</td>
<td>900</td>
<td>2100</td>
<td>rectangle</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>0000 to 1200 Jan.</td>
<td>975</td>
<td>1950</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>1200 to 1800 Jan.</td>
<td>750</td>
<td>1800</td>
<td></td>
</tr>
</tbody>
</table>

quantitative comparisons between averaged parameters (Table 2). Tests performed by varying the width of the frontal boxes did not show any qualitative changes.

2. STABILITY OF THE WESTERN PART OF THE PRIMARY FRONT

In this section, the stability of the primary front with respect to ambient strain is examined using the ARPEGE operational analyses every 6 hours from 1200 UTC 16 February to 1200 UTC 17 February (Exp. O, see Table 1). Following the work of Rivals et al. (1998), the stability is diagnosed by comparing the environmental along-front stretching in two different areas: the western part of the front (box \( B_W \)), where IOP17 cyclogenesis will develop; and the eastern part of the front (box \( B_E \)), closer to the parent low centre. During the considered time period, the western part of the front weakens and remains stationary (between 80°W and 60°W), whereas the more intense eastern part moves eastwards (Fig. 3). This clear contrast between the western and eastern parts of the front is the motivation for the positioning of frontal boxes \( B_W \) and \( B_E \).

(a) Environmental along-front stretching evolution

The evolution of the average along-front stretching of the environmental wind, \( \gamma_e \), is computed in frontal boxes \( B_W \) and \( B_E \), throughout the 24-hour period. Bishop (1996b) and Rivals et al. (1998) have shown that these averaged values are much less erratic than direct estimates made by using the observed wind. This better consistency was the main motivation for Bishop (1996a) in developing the domain-independent attribution technique. Values of \( \gamma_e \) obtained through the domain-independent attribution technique provide a clearer background signal which can be compared to its theoretical counterpart, the large-scale deformation used in idealized models. The theoretical critical threshold \( \gamma_c = 0.6 \times 10^{-5} \) s\(^{-1}\), defined by Bishop and Thorpe (1994b) as the upper strain limit allowing frontal wave development, has been confirmed by Chaboureau and Thorpe (1999) in a selection of FASTEX cases.

Table 3 presents the time evolution of \( \gamma_e \) computed in boxes \( B_W \) and \( B_E \) from 1200 UTC 16 February to 0600 UTC 17 February. For 1200 UTC 17 February, only the value in box \( B_W \) is shown, as box \( B_E \) has moved out of the analysis domain by this time. Values obtained for box \( B_W \) become appreciably less than \( \gamma_c \) after 1800 UTC 16 February, whereas values for box \( B_E \) always remain above it. Values within box \( B_W \) also show a decreasing trend until 0000 UTC 17 February; it may be conjectured that this decrease might correspond to the mature stage of the parent low. Box \( B_E \) shows
TABLE 3. Time evolution of the environmental averaged along-front stretching \( \gamma_e \) in the frontal boxes in Exp. O

<table>
<thead>
<tr>
<th>Time (UTC)</th>
<th>Box B_W</th>
<th>Box B_E</th>
</tr>
</thead>
<tbody>
<tr>
<td>1200 16 Feb</td>
<td>0.72</td>
<td>0.68</td>
</tr>
<tr>
<td>1800 16 Feb</td>
<td>0.44</td>
<td>1.14</td>
</tr>
<tr>
<td>0000 17 Feb</td>
<td>0.15</td>
<td>0.99</td>
</tr>
<tr>
<td>0600 17 Feb</td>
<td>0.21</td>
<td>0.79</td>
</tr>
<tr>
<td>1200 17 Feb</td>
<td>0.36</td>
<td></td>
</tr>
</tbody>
</table>

a similar evolution with a 6-hour delay, in agreement with the respective distances of box B_W and B_E from the parent low centre. So, this first result seems consistent with the work of Rivals et al. (1998, for their B1 and B2 boxes in their Fig. 12): both the timing and the location of the low-triggering seem to fit the environmental forcing. Later, after 0600 UTC 16 February, \( \gamma_e \) weakly increases again when the frontal cyclone starts developing.

(b) Potential for instability of the front

Latent-heat release in convective areas produces a dipole of PV sources giving a positive PV anomaly at low levels (Thorpe and Emanuel 1985). The ascending branch of active fronts is, in this way, often associated with a low-level strip of positive PV under the level of maximum condensation. As this strip of low-level PV anomaly satisfies the necessary condition for instability given by Charney and Stern (1962), it may lead to a linearly growing frontal wave instability, which was theoretically studied by Schär and Davies (1990), Joly and Thorpe (1990), Dritschel et al. (1991), and Bishop and Thorpe (1994a,b). Another possible growth mechanism is the finite-amplitude interaction of the low-level anomaly with an upper-level-anomaly, as considered by Thornicroft and Hoskins (1990). The evolution of the low-level PV field between 1800 UTC 16 February and 0600 UTC 17 February is investigated at the 700 hPa level. A strip of PV values higher than 0.5 PVU\(^*\) appears in the western part of the primary front (north and east of 70°W, 30°N) during this period (Fig. 5). On a vertical cross-section taken across the front (see Fig. 5(b) for location), PV values higher than 0.5 PVU can be seen up to the 550 hPa level (Fig. 6(a)), and the strip location corresponds to the moist frontal ascent branch (Fig. 6(b)). Cammas et al. (1999) further show that this frontal ascent coincides with the ascending branch of the TDA circulation near the upper-level jet entrance when the jet intensifies at 0000 UTC 17 February. Therefore, although the 950 hPa cyclonic vorticity decreases, the western part of the primary front displays an increasing potential for frontal instability at the 700 hPa level in the last 12 hours prior to the frontal cyclone triggering, because this area is exactly under the action of the intensifying jet TDA circulation (Fig. 4(b)).

(c) Triggering

In this subsection further results are presented using the 'vorticity wavyness' \( W(\xi) \), introduced by Renfrew et al. (1997). The authors define it as the peak vorticity, \( \xi_{\text{peak}} \), minus the maximum along-front average vorticity of the front, \( \max(\xi_y) \), in the rectangular box where the wave develops (box B_W in the present case), where \( y \) is

\[ 1 \text{ PVU} = 10^{-6} \text{ K m}^2\text{s}^{-1}\text{kg}^{-1} \] (Hoskins et al. 1985).
Figure 5. 700 hPa potential vorticity (contour interval 0.5 PVU, shaded above 0.5 PVU), overlaid with 700 hPa wind barbs (conventional plot in knots): (a) at 1800 UTC 16 February; (b) 0600 UTC 17 February, with the bold line marking the location of the vertical cross-section in Fig. 6.

Figure 6. Vertical cross-section in the entrance region of the upper-level jet at 0600 UTC 17 February (see Fig. 5(b) for location): (a) potential vorticity (solid lines, contour interval 0.25 PVU, shaded between 0.5 and 1 PVU) overlaid with isentropes (dashed lines, contour interval 3 K), (b) ageostrophic wind vectors (scale at bottom right), overlaid with normal wind component (solid lines, contour interval 5 m s\(^{-1}\)) and relative humidity (dashed lines, contour interval 25%, shaded above 75%).

The along-front direction. Renfrew et al. (1997) show that it can be used as an indirect measure of the nonlinear growth rate, and is suitable for work with real data because of low sensitivity to along-front changes. The time evolution of \(\max[\xi^2]\) (Fig. 7(a)) shows it decreasing until 0600 UTC 17 February, and then increasing rapidly after 1200 UTC 17 February, suggesting that the incipient cyclogenesis intensifies vorticity within box B\(\text{W}\). The time evolution of \(W(\xi)\) is fairly similar except that \(W(\xi)\) values increase 6 hours earlier. The evolutions of the previous vorticity-derived values confirm the time chosen to be that of cyclogenesis triggering. In the same way as Renfrew et al. (1997) and Chaboureau and Thorpe (1999), we plot the growth rate against the ambient normalized strain (Fig. 7(b)): the growth rate is measured as the rate of change
of vorticity waviness, and the ambient normalized strain as the along-front stretching normalized by the minimum strain rate $\alpha_{\text{min}}$ necessary to suppress barotropic frontal waves according to Bishop and Thorpe (1994b). Figure 7(b) allows us to check that the growth rate is increasing through the last 12 hours of the studied period, and that the wave is developing under normalized strain of less than 1, in agreement with the aforementioned studies.

To summarize this section: the strain analysis of the western part of the primary surface front shows (i) that the ambient along-front stretching is weak and subcritical in this area 12 hours prior to the IOP17 low-triggering (section (a)); and (ii) that the potential for frontal instability simultaneously increases, due to the intensification of ascent in the jet entrance (section (b)). The frontal cyclone seems to be triggered when and where both the environmental deformation and the potential for instability become favourable.

3. ROLE OF THE CONTINENTAL SURFACE LOW

We now turn to the examination of the remote action of the continental surface low. This discussion will use comparisons of the four datasets (the operational analysis and experiments A, B and C). These datasets were introduced in section 1, and are summarized in Table 1. The unmodified ARPEGE operational analysis used in section 2
is hereafter referred to as 'experiment O'. Experiments A, B and C are ARPEGE forecasts made from modified initial fields obtained through PV inversion performed by Arbogast and Joly (1998b). For each of these four datasets we will compute the environmental contributions to the wind fields and, when required, perform a frontal-strain analysis using the tools of the previous sections.

(a) Unmodified ARPEGE operational analysis

We begin with operational analyses of experiment O. In order to study the role of the continental surface low, we now use the domain-independent attribution tool to repeat the strain analysis of the western part of the primary front (box B_W) presented in section 2(a), but this time we use a modified wind field excluding the contributions from the surface continental low. In other words, this means that we compute an environmental wind field in which the vorticity/divergence elements contained in both boxes* B1 (collocated with the continental surface low) and B_W (encompassing the western part of the primary front) are set to zero.

The time evolutions of the along-front stretching, \( \gamma_e \), averaged over box B_W when the surface continental low is accounted for (solid line, computed in section 2(a)), and without this low (dotted line) are compared in Fig. 8(a). The shapes of the two curves are similar, but discarding the surface continental low increases the environmental strain by 5% to 30%. However, both lines show stretching values lower than the Bishop and Thorpe (1994b) critical threshold, \( \gamma_c \), after 1800 UTC 16 February. The impact of the surface continental low is larger between 0000 UTC and 0600 UTC on 17 February. At 1200 UTC, IOP17 cyclogenesis is already well under way, and the remote impact of the continental surface becomes negligible. From this instantaneous diagnosis, the main effect of the surface continental low seems to be to weaken the along-front stretching in the western part of the primary front just before onset of the frontal cyclone.

(b) ARPEGE forecasts

Experiments A, B and C (Table 1) are now used to examine how the impact of the surface continental low builds up as a function of time. First, experiment A is initialized with the unmodified 1200 UTC 16 February analysis, and the ARPEGE model is run to provide 6-hourly wind fields. The frontal cyclone development is correctly captured, giving a 1200 UTC 17 February forecast surface-pressure minimum of 1022 hPa, to be compared with the 1018 hPa minimum found in the analysis. Once more, we repeat the strain analysis for the western part of the primary front, with a B_W box slightly shifted to stay phase-locked with the forecast frontal-surface vorticity maxima (Fig. 9(a)). The time evolution of \( \gamma_e \) in experiment A is shown as a solid line in Fig. 8(b). The overall strain evolution is similar to the one shown for experiment O (operational analysis) in Fig. 8(a), with a minimum value found at 0000 UTC 17 February. However, the range in strain variation forecast in experiment A is smaller than the range found in the operational analysis. This underestimation is probably due to the coarser resolution of the model in experiment A, compared to the one in experiment O (Table 1).

In experiment C, the ARPEGE run also starts at 1200 UTC 16 February, but is initialized with modified fields obtained by removing the PV secondary maximum corresponding to the continental surface low. In this case the frontal cyclone cyclogenesis is delayed (triggering at 0000 UTC 18 February) and much weaker (Arbogast and Joly 1998b). Again \( \gamma_e \) is computed for a box B_W slightly shifted to fit the forecast low-level vorticity maxima, as above. The \( \gamma_e \) time evolution is shown by the dashed line in

* Box locations are shown in Fig. 3, and detailed in Table 2.
Fig. 8(b). The overall trend is similar to experiment A, but the along-front stretching in experiment C is always larger by 20% to 30% until the end of the run. This confirms that the continental surface low acts to decrease the along-front stretching, and that its time-integrated effect is noticeable for the whole 24-hour time period considered. However, Fig. 9(c) also suggests that at 0600 UTC on 17 February a weak cyclonic low-level circulation (near 37°N, 75°W) might already be regenerated by the upper-level anomaly.

Finally, in experiment B the tropopause-level anomaly embedded within the upper-level trough over the continental USA (located over Lake Superior on Fig. 1(a)) is considered as another possible candidate to interact with the IOP17 frontal cyclone. Hence, experiment B is performed by removing this upper-level cyclonic PV anomaly from ARPEGE initial fields at 1200 UTC 16 February, and running the model. In this case, the direct impact is on the continental surface low itself, which is substantially weakening with time. In Fig. 9(b) the low-level relative vorticity signature of the continental surface low in experiment B (40°N, 80°W) is even weaker than the residual low noticed near 37°N, 75°W in experiment C (Fig. 9(c)). The time evolution of $\gamma_e$ in box $B_w$, computed as before, is shown by the dash-dotted line on Fig. 8(b). The $\gamma_e$ value in experiment B is half-way between those in experiment A and C, with a value very close to experiment A (forecast from unmodified analysis) at 1800 UTC 16 February and, by contrast, close to experiment C (no continental low initially) by 0000 UTC 17 February. Thus, our conclusion for the present section is twofold: (i) the strength
of the surface-level continental low seems to be linked to the upper-level PV anomaly; (ii) the environmental along-front stretching in box $B_W$ is controlled by the intensity of the continental surface low—the stronger the continental surface low, the weaker the environmental along-front stretching in box $B_W$.

4. MECHANISMS FOR THE REMOTE ACTION OF THE SURFACE CONTINENTAL LOW

In the previous sections, we have seen that the IOP17 frontal cyclone is triggered in a low-strain area of the western part of the primary surface front, and that the environmental strain in this area is influenced by the intensity of the surface continental low. Our goal is now to examine different mechanisms by which the continental surface low might modify the baroclinic zone where the IOP17 frontal cyclone develops. For this purpose, the attribution technique is used in connection with a different wind field partition. We consider box $B_L$, chosen to encompass the continental surface low (Table 2 and Fig. 3), and define the wind contribution attributed to the continental surface low as
the wind field computed from the vorticity/divergence elements contained inside the box \( B_1 \). We will focus on how the along-front stretching and frontogenesis inside box \( B_W \) are modified by this flow, and also discuss the contribution of the flow of the continental low to the thermal advection in the IOP17 frontal cyclone development zone.

\[(a) \text{ Wind component attributed to the surface continental low}\]

As we mentioned earlier (section 1(b)), the continental surface low is the sole coherent dynamical vorticity and divergence structure found in the vicinity of the western part of the primary front by the time of the frontal cyclone onset. So, between 1200 UTC 16 February and 0600 UTC 17 February we define box \( B_1 \) as a constant-area rectangular box, phase-locked to the vorticity/divergence elements associated with the continental surface low and with its embedded fronts (Table 2 and Fig. 3). At 1200 UTC 17 February, the continental surface low moved so close to the primary front that we had to make box \( B_1 \) slightly smaller, to avoid intersecting box \( B_W \). It is interesting to note that the cold front of the continental low is more clearly delineated than its warm front as long as the low moves over land, whereas the warm front becomes stronger than the cold front when the low moves offshore. This difference between continental and marine cyclones was noticed for instance by Kuo and Low-Nam (1994) in a numerical investigation of the sensitivity of cyclogenesis to surface friction.

Figure 10(a) shows the cyclonic circulation attributed to the continental surface low at 0000 UTC 17 February, i.e. the winds \( U_{i\phi} \) computed from the vorticity elements contained by box \( B_1 \). The divergent flow attributed to the continental low, \( U_{ix} \), computed from the divergence elements contained by box \( B_1 \), clearly depicts the convergence towards the centre of the surface low and along its associated fronts (Fig. 10(b)). The modulus of \( U_{i\phi} \) is roughly twice the modulus of \( U_{ix} \), with values rapidly decreasing with radial distance from the centre. The resulting total circulation attributed to the continental low \( (U_i = U_{i\phi} + U_{ix}) \) is southerly within box \( B_W \) in the western part of the primary front (Fig. 10(c)). The magnitude of this wind \( U_i \) (about 6 m s\(^{-1}\)) is 30% of the analysed total wind magnitude at the same location. As the intensity of the vorticity/divergence elements inside box \( B_1 \) remains nearly constant, the symmetrical shapes of the \( U_{i\phi}, U_{ix} \), and total wind \( U_i \) contours remain almost unchanged between 1200 UTC 16 February and 0600 UTC 17 February (not shown). However, as the continental low centre moves south-eastwards, its contribution \( U_i \) to the total wind in the western part of the primary front (box \( B_W \)) increases somewhat with time.

\[(b) \text{ Stretching and frontogenesis}\]

Another factor influencing frontal wave development is frontogenesis. Bishop (1993b) and Parker (1998) comment on the fact that frontolytic deformation can increase the growth rate of instabilities, and Ayralut et al. (1995) mention that their ‘type 2’ cyclones develop better when the background flow is frontolytic. Following Bishop (1996b) and Rivals et al. (1998), we use deformation fields and Keyser et al. (1988) frontogenesis vectors to quantify the frontogenetic forcing in box \( B_W \). These diagnostics are computed from the total analysed wind field and the wind field attributed to the continental low, at 0000 UTC 17 February when the instantaneous remote effect of the continental low is the greatest in terms of along-front stretching (section 3(a)).

Figure 11(a) depicts the total-deformation field associated with the unmodified analysed wind by means of double arrows. The orientation of the double arrows gives the orientation of the dilatation axis of the total deformation, and the length of the arrows indicates the magnitude of the strain rate (half that of the total deformation).
Figure 10. Top panels: vectors (scale at bottom right) and modulus (every 2 m s$^{-1}$) of: (a) wind $U_{v_e}$ computed from vorticity elements isolated in box $B_1$; (b) wind $U_{e_d}$ computed from divergence elements of box $B_1$. (c) Wind vectors ($U_{v_e} + U_{e_d}$) attributed to the continental low, in box $B_W$ (location given in (a) and (b)). All for 0000 UTC 17 February.

The orientation of the dilatation axis in box $B_W$ makes an angle of roughly 45° with the isotherms, which implies a rotation of the thermal gradient without significantly changing its intensity. Figure 11(b) shows the total-deformation arrows computed with the $U_i$ winds, associated with the continental low. In this case the dilatation axis is orientated almost perfectly across the isotherms, and therefore acts to decrease the thermal gradient.

Figure 11(c) depicts the frontogenesis vectors computed from the unmodified analysed winds. These vectors point towards warm air, indicating a frontogenetic configuration in box $B_W$ (Keyser et al. 1988). Conversely, the frontogenesis vectors computed from $U_i$ winds (Fig. 11(d)) point in the opposite direction, towards cold air. The continental-low contribution is therefore frontolytic in box $B_W$. Note that it counteracts almost 20% of the total wind effect in terms of across-front component magnitude.

(c) Thermal advection

Finally this subsection examines the thermal advections computed, as above, either directly from the unmodified ARPEGE analysed winds, or from $U_i$ winds attributable to the surface continental low. Very recently, Chaboureau and Thorpe (1999) have proposed a more general approach which extends the wind partitioning of Bishop (1996a) to the temperature field by assuming a balance condition. It is important to note that the approach used here is simpler (and maybe coarser) as the thermal advection computed from the $U_i$ winds corresponds to the transport of the total temperature fields by winds attributed to the continental low.
Figures 12(a) and 12(b) show the thermal advection by the total analysed winds, and by the $\mathbf{u}_i$ winds at 0000 UTC 17 February, respectively. The signature of the primary front is clearly depicted by the sign reversal of the total thermal advection (Fig. 12(a)). The transport of the total temperature field by the wind attributed to the low (Fig. 12(b)) also clearly depicts the fronts associated with the continental surface low (north-west of it). However, the shape is different to the south-west, with a band of positive values ($5$ K day$^{-1}$) reaching the western part of the primary front. This warm advection persists during the whole period considered (1200 UTC 16 February to 1200 UTC 17 February, not shown); it acts on both sides of the primary front, with a near cancellation of the post-frontal cold advection (42°N, 73°W), and an increase of the warm-sector warm advection (by 20% to 25%). The thermal advection associated
Figure 12. 950 hPa thermal advection (dashed for negative values), overlaid with wind vectors (scale at bottom right), computed (a) with the total wind (contour interval 10 K day\(^{-1}\)), (b) with the wind U, attributed to the continental low (contour interval 5 K day\(^{-1}\)), at 0000 UTC 17 February.

with the continental surface low, therefore, seems to reduce the primary front thermal contrast. This conclusion confirms the overall frontolytic pattern found in the previous subsection. Another (albeit somewhat conventional) remark can be made if we interpret the warm advection in the framework of the classical omega equation: in the warm sector we expect low-level warm thermal advection to be associated with mid-tropospheric rising motion and cyclonic evolution. Moreover, frontal wave theories of Schär and Davies (1990) and Joly and Thorpe (1990) also suggest that a low-level warm anomaly south of the primary front favours development.

To summarize, it has been shown in this section that the wind associated with the continental surface low weakens the along-front stretching through a frontolytic effect, and favours a cyclogenetic evolution by increasing the advection of warm air south of the primary front.

5. CONCLUSION

The role of a pre-existing continental surface low on frontal cyclone triggering during FASTEX IOP17 has been investigated. In a first step, the stability of the primary cold front on which the frontal cyclone develops was examined in terms of along-front stretching using the domain-independent vorticity and divergence attribution technique of Bishop (1996a,b) on ARPEGE analyses. The western part of the primary front, where the frontal cyclone finally develops, exhibits a decreasing along-front stretching, and becomes smaller than the Bishop and Thorpe (1994b) critical threshold 12 hours before the frontal wave onset. It is shown that during the triggering period the growth rate is increasing and that the wave is developing under normalized strain less than 1, in agreement with studies by Renfrew et al. (1997) and Chaboureau and Thorpe (1999). During the same period, the potential for instability of the western part of the primary front increases, as this area is located under the ascending branch of the transverse and direct ageostrophic circulation at the entrance of an upper-level jet streak. In a second step, the remote action of the surface continental low on the primary front was investigated by partitioning the wind field into contributions from the environment and
from the low. Comparing the results found using both unmodified analyses and three ARPEGE forecasts conducted by Arbogast and Joly (1998b), it was shown: (i) that the continental surface low weakens the along-front stretching by roughly 5% to 30% in the western part of the primary front; (ii) that the intensity of the continental surface low itself seems linked to an upper-level PV anomaly further west. Finally, a brief discussion was put forward of the possible mechanisms by which the continental surface low might modify the baroclinic area where the IOP17 frontal cyclone develops. In addition to the environmental-strain modification mentioned above, the results show that the continental surface low has a frontolytic effect on the primary front, and is also associated with a thermal advection pattern promoting rising motion (thereby favouring diabatic processes) in the warm sector of the incipient frontal cyclone. Our conclusion, therefore, suggests that the triggering of the IOP17 frontal cyclone fits well with the theoretical framework of Bishop and Thorpe (1994a,b), albeit a more complex precursor pair is found, but that the further development of the wave might rely more on a delayed type B development process with the upper-level disturbance, as suggested by Arbogast and Joly (1998b) and Cammas et al. (1999), and on diabatic processes favoured by the thermal advection patterns.

ACKNOWLEDGEMENTS

The authors are particularly grateful to Alain Joly for his invaluable and enthusiastic support of FASTEX, and to Keith Browning who chaired the FASTEX Operation Committee in Shannon. We also express our appreciation to the NOAA Gulfstream IV flight crews and engineers for the dedication in making difficult observations, and to all the individuals and organizations who made the execution of FASTEX possible. We also thank Daniel Ramond and two anonymous referees for their very helpful criticisms. FASTEX has been supported by INSU/PATOM under contract 97/01, and by the European Commission under contract ENV4-CT96-0322. Météo-France is acknowledged for free access to SYNERGIE. Computer resources were allotted by IDRIS (projects 97569, 98569, and 981076), and by Météo-France. This work represents a portion of the first author’s thesis dissertation, and was supported by a Météo-France FCPLR academic leave.

APPENDIX A

The PV-inversion framework

The design of a quasi-geostrophic inversion method (Pedlosky 1987; Hoskins et al. 1985) is presented here. It is applied to real cases to demonstrate which PV anomalies, present within the atmosphere prior to the development, are involved in the cyclogenesis. The strategy, based upon the PV perspective is outlined here; it involves four steps (Arbogast 1998).

1. All ‘coarse grain’ PV elements are identified.
2. For each PV element, the ‘balanced’ part of the flow is deduced using a variational implementation of PV inversion (Arbogast and Joly 1998a).
3. The ‘balanced’ part of the flow corresponding to the PV element of interest is removed from the initial conditions of the ARPEGE model.
4. A forecast with the revised fields is then performed.
The system of partial derivative equations corresponding to the invertibility principle compatible with ARPEGE variable (see Courtier et al. (1991) for more details) involved in this paper takes the following form:

\[
\text{Quasi-geostrophic PV (QGPV): } \quad q_g - f = \nabla^2 \Psi + f_0 \frac{\partial}{\partial p} \left( \Theta \frac{\partial T_r}{\partial p} \right)
\]

balance: \[ \nabla^2 (\Phi + R T_r \ln p_s) - \nabla \cdot f \nabla \Psi = 0 \]

hydrostatic: \[ \frac{\partial \Phi}{\partial p} = \frac{R}{p} \left( \frac{p}{p_0} \right)^{R/C_p} \Theta, \quad (A.1) \]

where \( q_g \) is the quasi-geostrophic pseudo-PV, \( \Psi \) is the stream function, \( p \) is the pressure, \( R \) is the universal gas constant, \( T_r \) is the ARPEGE isothermal reference-state temperature, \( C_p \) is the specific heat of air at constant pressure, \( \Theta \) is the potential temperature, \( \Theta_r \) is the potential temperature of the quasi-geostrophic reference state, \( \Phi \) is the geopotential, \( p_s \) is the surface pressure, \( f \) is the Coriolis parameter and \( f_0 \) its value taken at a reference latitude. Since the quasi-geostrophic system on the sphere used by Simmons and Hoskins (1976) does not possess any obvious Lagrangian invariant, we chose the very common form of QGPV on the \( \beta \)-plane. Unlike QGPV, the present balance equation is relevant to the sphere. It is the zero-order assumption of the divergence equation of ARPEGE. The \( R T_r \ln p_s \) term is due to the generalized vertical coordinate, whereas no simplification of the Coriolis parameter is assumed. If \( p_s \) and either \( \Theta \) or \( \Psi \) are known the solution of (A.1) is unique.

Equation (A.1) is solved using the variational method introduced by Arbogast and Joly (1998a). Let \( \mathcal{P} \) be the discretized analogue of the right-hand side of the first equation of (A.1). \( \mathcal{P} \) applies on the non-divergent variable \( X \) of ARPEGE, namely (\( \Phi, T_r \)). \( \mathcal{B} \) is the discretized form of the departure from balance. The solution of (A.1) is also the minimum of the function:

\[
J = \frac{1}{2} (\mathcal{P} X - \delta q_g)^*(\mathcal{P} X - \delta q_g) + \frac{1}{2} X^* \mathcal{B}^* \mathcal{B} X, \quad (A.2)
\]

where \( \delta q_g \) is a known QGPV field and \((\cdot)^*\) stands for transposition. An approximate solution can be found using a quasi-Newton based method (Gilbert and Lemaréchal 1989).

For many reasons, it is not useful to deal with the state variable of the model in full vertical and horizontal resolution. The first reason is that the quasi-geostrophic assumption is intrinsically only relevant to large-scale flows. The second reason is related to the fact that only the coarse-grain QGPV anomalies are the key features for mid-latitude cyclogenesis. For these reasons we choose a T63 spectral resolution on the unstretched sphere and 19 vertical levels. The corresponding horizontal resolution in physical space (a few hundred km) is suitable to be compared with the density of upper-air measurements used in the global model analyses.

For experiments B and C, the PV initial conditions modification by inversion was used in a systematic way. The contribution of each PV element, at upper and lower levels, has been singled out with the ARPEGE model. The anomalies depicted in experiments B and C were the only ones which contribute significantly to the triggering of the IOP17 cyclone. In experiment B, where an upper-level feature is removed, homogeneous Dirichlet boundary conditions are used. That means that no wind anomaly at the first level, just above the ground, is attributable to the upper-level PV anomaly. In experiment C, where a low-level feature is removed, non-homogeneous Dirichlet
boundary conditions are used. The low-level PV anomaly as well as the surface vorticity anomaly are inverted.

**APPENDIX B**

*The domain-independent attribution method*

The domain-independent vorticity and divergence attribution technique pioneered by Bishop (1996a,b) allows the partitioning of the wind field into a frontal part and an ambient part, by defining a ‘frontal box’ enclosing the surface frontal vorticity strip. Readers are referred to the original papers for the technical details of the method, only a summary is given here.

The partitioning of a horizontal wind field into irrotational and nondivergent parts, on a finite domain, is achieved by solving the Poisson equations, for the stream function from vorticity and the velocity potential from divergence, using free-space Green’s functions. These remove the boundary conditions to infinity, and so give a certain domain independence to the stream function and velocity-potential solutions. Each grid point of the domain is associated with a disc of uniform vorticity and a disc of uniform divergence, calculated from circulation and flux estimates of the neighbouring grid boxes. Each element of vorticity and divergence is therefore associated with an unique wind field, obtained via the Poisson equation solutions. The flow obtained by summing the winds attributable to all the elements of vorticity in the finite domain defines the total rotational flow, \( U_\psi \). Similarly, summing the winds attributable to all the elements of divergence in the finite domain gives the divergent flow, \( U_\theta \). The ‘remainder’ harmonic part of the wind, \( U_\theta \), is both solenoidal and irrotational in the finite domain, and is due to vorticity and divergence outside the domain. This three-component partition of the observed wind \( (U_{\text{obs}} = U_\psi + U_\chi + U_\theta) \) is unique and independent of the domain boundaries.

Bishop (1996b) proposes a piecewise-like methodology to partition the observed flow into a frontal part and an environmental part, by defining a frontal box around a limited part of the surface front. The frontal flow \( (U_f) \) is defined as the wind attributable to both the vorticity elements \( U_{\psi f} \) and the divergence elements \( U_{\chi f} \) inside the frontal box. The environmental flow is the external flow \( (U_e) \) associated with vorticity elements \( U_{\psi e} \) and divergence elements \( U_{\chi e} \) external to the frontal box, and vorticity and divergence elements outside the finite domain \( (U_\theta) \); so that \( U_{\text{obs}} = U_e + U_f = (U_{\psi e} + U_{\psi f}) + U_{\chi e} + U_{\chi f} + U_{\theta} \). All computations of wind derivatives are done in a frontal coordinate system. The long sides of the box are in the along-front direction, and for a cold front the across-front direction points towards the warm air.

**REFERENCES**


Kuo, Y.-H. and Low-Nam, S. 1994  ‘Effects of surface friction on the thermal structure of an extratropical cyclone’. In Proceedings of the conference on the life cycles of extratropical cyclones, Bergen, Norway, 27 June to 1 July. Available from the University of Bergen, Norway
<table>
<thead>
<tr>
<th>Author(s)</th>
<th>Year</th>
<th>Reference</th>
</tr>
</thead>
<tbody>
<tr>
<td>Pedlosky, J.</td>
<td>1987</td>
<td><em>Geophysical Fluid Dynamics, 2nd edition</em>. Springer-Verlag, New York, USA</td>
</tr>
<tr>
<td>Renfrew, I. A.</td>
<td>1995</td>
<td>The development of secondary frontal cyclones. PhD dissertation, University of Reading (Available from the Department of Meteorology, University of Reading, PO Box 243, Reading RG6 6BB, UK)</td>
</tr>
</tbody>
</table>