Dynamical structure of a wide cold-frontal cloudband observed during FRONTS 87

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SUMMARY

Conditional symmetric instability (CSI) is examined as a possible mechanism for generating cloudbands within a cold-frontal system. The data used in this study were collected during the Franco-British field experiment FRONTS 87. A dense network of rawinsonde measurements was used to test the various hypotheses underlying CSI, such as the sign of the moist potential vorticity, the relative slope of absolute momentum and moist isentropic surfaces or the alignment of bands with the vertical shear of the wind inside the unstable domain. Radar data provided from the French dual-Doppler radar system RONSARD was used to retrieve the three-dimensional airflow within a cloudband observed during the last frontal event of FRONTS 87. A new retrieval method, called MANDOP, was used to estimate the circulations within this band and to test whether the retrieved circulations were consistent with the CSI theoretical expectations. A roll-like circulation present in the vicinity of the observed cloudband seems to support the hypothesis that CSI was the formative mechanism.

1. INTRODUCTION

The banded structure of precipitation has been the subject of considerable observational and theoretical study over the last 25 years. Improvements in mesoscale observational technology such as the development of meteorological radars and satellite imagery have provided the observational evidence of this banded structure, and several papers (Browning and Harrold 1969; Houze et al. 1976; Matejka et al. 1980; Parsons and Hobbs 1983; Lemaitre et al. 1989) have described various aspects of the observed rainbands. Several observational studies have attempted to demonstrate the possible role of conditional symmetric instability in the formation of precipitation bands (Bennetts and Sharp 1982; Bennetts and Ryder 1984; Seltzer et al. 1985; Wolfsberg et al. 1986; Lemaitre and Scialom 1992a,b; Lemaitre and Testud 1988). In parallel, numerical models (Knight and Hobbs 1988), have shown that CSI may play an important role in the formation and intensification of banded structure within a convectively stable region and that many of the characteristics of these numerically modelled bands agree with the theory of CSI.

Our purpose has been to consider the relevance of CSI to a wide cold cloudband formed within an atmospheric front. The relevant data were collected during the cooperative Franco-British field experiment FRONTS 87, which took place during the autumn and winter of 1987-88. A primary scientific objective of the experiment was to study the generation of mesoscale rainbands associated with a cold front and to test theories and models of CSI.

In this paper we report the results of an analysis of a cold front observed during 12 January 1988 (intensive observational period no. 8—hereafter IOP 8). First, the vertical structure of the front is described, with the aid of time-height cross-sections of various thermodynamic and dynamic parameters constructed from data provided from upper-air sounding stations. The frontogenetic forcing and ageostrophic circulation associated with this case have not been considered here in detail; that has been done by Lagouvardos et al. (1992). For this analysis the velocity azimuth display (VAD) sequences from the C-band dual-Doppler radars of the RONSARD system were used when deducing the three-

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dimensional wind circulation, within the observed cloudband, by a new method called MANDOP analysis (Scialom and Lemaitre 1990) which was used to retrieve the three components of the wind within the banded region of precipitation. Circulations associated with CSI are often masked by other motions (mainly upright convection and ageostrophic circulation around the front), but this particular analysis allows us to isolate the circulation corresponding to symmetric overturnings and thus to test the hypothesis that the derived circulation is consistent with CSI theoretical expectations, such as the presence of a roll-like circulation, ascending motions sloping nearly along the moist isentropic surfaces and the buckling of absolute momentum surfaces (Bennetts and Hoskins 1979; Thorpe and Rotunno 1989).

In section 2, a brief review of the FRONTS 87 experiment is presented. Section 3 contains a summary of the theoretical basis of CSI and a review of observational evidence of CSI. The cold front of 12 January 1988 and its associated banded organization are described in section 4. Section 5 considers the possible role of CSI by way of an analysis of rawinsonde measurements. The three-dimensional circulation obtained with the MANDOP method is discussed in section 6, and a conclusion is given in section 7.

2. The FRONTS 87 Experiment

The mesoscale frontal dynamic project (FRONTS 87) was a European experiment, designed to study the structure, evolution and dynamics of active cold fronts observed over an area centred on the Channel between England and France (Clough 1987). The Channel area offers a natural setting for such an experiment, being quite well served by the combination of British and French routine observing networks.

The experimental area consisted of a threefold nested structure, comprising a zone of intensive small-scale measurements (100 × 100 km²), an inner area (500 × 500 km²) and an outer observing area (1200 × 1200 km²). In this study, our interest is focused on the inner area over Brittany, France (Fig. 1), where the RONSARD system and an enhanced rawinsonde network were installed. In the French part of the experiment, three mobile sounding stations at Brest, Lannion and Lorient were operating during eight intensive observational periods. Their relative positions were chosen so as to form a nearly equilateral triangle with 80 km sides, thus permitting an accurate evaluation to be made of cross-front and along-front gradients within the triangle. This sampling pattern was complemented by a network of three VHF wind profilers (3 stratospheric/ tropospheric radars), three acoustic sounders, a UHF wind profiler at Brest, and a network of 10 ground-measurement stations dispersed within the triangle. The 5 cm (C-band) dual-Doppler radar system used in this study was located west of Brest (R1 and R2 in Fig. 1). The two radars were separated by a distance of 21 km along a line oriented 36° clockwise from north.

3. Review of Conditional Symmetric Instability

(a) Theoretical review

Conditional symmetric instability (CSI) was proposed by Bennetts and Hoskins (1979) as a possible explanation for frontal rainbands. CSI is a form of inertial instability along the wet-bulb potential-temperature (θₑ) surfaces, that can occur in a baroclinic flow generated from an unstable balance of pressure gradient, buoyancy and Coriolis forces. This type of instability manifests itself by rolls oriented approximately along the vertical shear of the horizontal wind vector (the 'thermal wind').
A key feature of this theory is the sign of the potential vorticity. In the context of a quasi-two-dimensional front, the instability grows as soon as the moist potential vorticity, $q_w$, defined as

$$q_w = \frac{fg}{\theta_0 \rho} \mathbf{z}_a \cdot \nabla \theta_w$$

becomes negative (here $f$ is the Coriolis parameter, $g$ the gravitational acceleration, $\theta_0$ a reference value of potential temperature, $\mathbf{z}_a$ the absolute vorticity, $\theta_w$ the wet-bulb potential temperature and $\rho$ the air density). As the instability grows, differential advection overturns the layers of the mid troposphere, generating classical gravitational instability. The resulting convection leads to a banded organization of the precipitation.

For a two-dimensional flow, the aforementioned criterion can be expressed as a function of the relative slope of a wet-bulb potential-temperature surface and an absolute-momentum surface, $M = v + fx$, where $v$ is the wind component in the thermal wind direction and $x$ is the coordinate perpendicular to it. In particular, if the moist isentropic surfaces are more vertical than the $M$-surfaces, the atmosphere is symmetrically unstable. This type of instability is frequently referred to as 'slantwise convection', since the motion is confined between the $M$- and $\theta_w$-surfaces which are in general tilted relative to the vertical in the $x$-$z$ plane. This criterion has been widely used to identify a region where CSI can develop (Emanuel 1983, 1988; Sanders and Bosart 1985) and has the great advantage that it involves only primary observed quantities (vertical profiles of wind, temperature and humidity) which can be obtained easily from classical rawinsoundings. For the purpose of this study, both criteria (sign of moist potential vorticity and relative slope of $M$- and $\theta_w$-surfaces) have been used, since their methods of evaluation are essentially independent (calculation of gradients with a centred difference scheme and comparison of vertical profiles of $\theta_w$ and $M$, respectively).
According to the theory of symmetric instability, the perturbation growth is derived primarily from the kinetic energy of the mean flow. As pointed out by Bennetts and Hoskins (1979), the time rate of change of the eddy kinetic energy, $E'_e$, of the most unstable mode (ascending and descending motions along isentropes) can be written as

$$\frac{\partial E'_e}{\partial t} = -u'v' \frac{\partial V}{\partial x} - v'w' \frac{\partial V}{\partial z}$$

(2)

where $u'$, $v'$ and $w'$ are perturbations of the three wind components and $V$ is the component of the basic flow in the thermal wind direction. Numerical simulations of CSI (Bennetts and Hoskins 1979; Thorpe and Rotunno 1989; Ducrocq 1989) have demonstrated that the main supply of energy for the perturbed circulation is the term involving $v'w'$ (term II), which is typically one order of magnitude larger than term I. This results from the fact that the horizontal gradient of the wind parallel to the front is much less than the vertical gradient (except in the vicinity of the frontal discontinuity near the surface). Taking this result into account, term I vanishes, and it follows from Eq. (2) that the perturbed circulation intensity increases mainly by a transfer of kinetic energy from the basic state when the basic vertical wind shear $\partial V/\partial z$ and the product of the $v'$ (along-front wind perturbation) and $w'$ (vertical velocity perturbation) have opposite signs. Following the sign of the basic vertical wind shear (positive/negative), the correlation between the $v'$ and $w'$ fluctuations (negative/positive) should be present within the atmospheric region where CSI is expected to develop (Bennetts and Hoskins 1979; Lemaitre and Scialom 1992, and personal communication; Lemaitre and Testud 1988).

Recently, studies have been carried out by Emanuel (1985) and by Thorpe and Emanuel (1985) dealing with the interaction between frontogenesis and a region nearly neutral to slantwise convection. In the latter paper, numerical calculations using a time-dependent semigeostrophic model show that the frontogenesis mechanism can account for the upward motion ahead of the frontal surface, but the existence of near neutrality to slantwise convection reduces the width of the ascending branch of the ageostrophic circulation, forming a narrow precipitation band. The authors hypothesized that frontal-related circulation produced by the geostrophic forcing is similar to slantwise convection, making it difficult to distinguish frontal from slantwise convection circulations in observational studies.

(b) Observational evidence of CSI

During the last decade several papers have examined the relevance of CSI in cloud and/or precipitating bands. These papers confirmed that the width and spacing of rainbands are mesoscale in nature and are oriented (more or less) along the thermal wind inside the unstable layer. It has also been verified that CSI-related bands have no propagation with respect to the mean flow within which they are embedded. However, Seltzer et al. (1985) reported that in some cases there exists a propagation of bands relative to the mean flow: a fact that classical CSI theory cannot explain. This would occur if the basic wind field is not in thermal wind balance as required by CSI theory. However, as the mean flow within the unstable layer is usually defined by soundings in the vicinity of the rainbands, circulations related to the CSI rolls may influence the wind vertical profile provided from the soundings. This point will be taken up in section 5 where wind hodographs will be presented.

The most widely used criterion for assessing symmetric instability is the comparison
of the absolute momentum, \( M \), and the equivalent potential temperature, \( \theta_e \) (or wet-bulb potential temperature \( \theta_w \)) relative slope. In parallel, several symmetric instability parameters are calculated (equivalent or wet-bulb potential vorticity, dry and moist Richardson numbers, slantwise convective available potential energy, perturbation growth rate) from individual soundings or from cross-sections when dense measurements are available. A review of 13 observational studies of CSI published after 1979 is given in Table 1.

4. The case-study of 12-13 January 1988 (IOP 8)

The mesoscale cloudband which was the object of our study was embedded in a cold-frontal system observed over Brest, France at 2330 GMT 12 January (hereafter referred to as time T). The surface map and 1000–500 hPa thickness patterns at 00 GMT 13 January (T + 0.5 h) are shown in Fig. 2. The cold front was aligned south-west–north-east and was associated with two low centres of 965 hPa situated north-west of Scotland (L1 and L2). The approximated frontal speed was estimated as being 6 m s\(^{-1}\), coming from the 300° sector. Synoptic observations at sea and overland recorded strong south-westerly winds ahead of the frontal surface. The 1000–500 hPa thickness pattern shows a relatively weak cross-front thermal gradient. The FRONT 87 ground stations in Brittany, recorded a strong veering of the wind and a sharp decrease of its magnitude at the moment of the frontal passage. Heavy rain accompanied this front with a maximum of 12 mm h\(^{-1}\) recorded at 2330 GMT at Brest. A well-marked pressure kick was also registered. In contrast, surface-level temperature fell only slightly (about 1 degC).

(a) Satellite imagery

An infrared image taken at the moment of the frontal passage over Brest by the geostationary satellite METEOSAT gives an overall view of the cloud system associated with this frontal zone (Fig. 3). The location of the two associated low-pressure centres over the North Atlantic Ocean are marked by letters L1 and L2. The warm conveyor belt, extending from southern Spain into northern Europe, occupied the major part of the cloud system covering the western sector of France. The most prominent feature of this belt was band B which had a width of approximately 40 km. It was located at the rear side of the surface cold front (shown with conventional symbols) but above the frontal zone, and oriented 40° clockwise from north. Successive infrared images taken every thirty minutes (not shown) suggested a propagation velocity of 26–28 m s\(^{-1}\) in a north-east direction (40–45° clockwise from north). The cloudband associated with the narrow cold-frontal rainband was at that time over Brest. Since its altitude did not exceed 3 km, it does not appear clearly on the satellite picture because it is embedded deep within the mass of stratiform cloud. Indeed, the mean altitude of the cloud cover seen in Fig. 3 was actually higher than 3 km. At least two other major rainbands were embedded within the warm conveyor belt ahead of the frontal system. They were observed over Spain, but since they were outside the effective range of the radar (100 km, denoted by a circle) they will not be discussed further. Apart from the banded organization another remarkable feature is the generation of a frontal wave in the time interval between time T – 5 h (image not shown) and time T. Its wavelength is about 1500 km. Band B later dissipated and five hours after the frontal passage over Brest it was no longer detectable in the infrared satellite pictures.

(b) Synoptic context

A description of the synoptic context of this particular case-study and the associated ageostrophic circulations can be found in the thesis by Lagouvardos (1992) and
<table>
<thead>
<tr>
<th>Source</th>
<th>Number of case-studies (c-s) and rainbands</th>
<th>Spacing between bands</th>
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<th>Absence of band propagation with respect to the mean flow</th>
<th>$\theta_v$ or $\theta_w$ surfaces steeper than absolute momentum $M$</th>
<th>Negative correlation between $\theta'$ and $w'$</th>
<th>Sign of symmetric instability parameters depending on $\theta_v$ or calculation of SCAPE$^1$</th>
<th>General remarks</th>
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<tbody>
<tr>
<td>Bennetts and Sharp (1982)</td>
<td>45 cases of banded precipitation one c-s several squall lines ahead of a cold front</td>
<td>nc</td>
<td>nc</td>
<td>nc</td>
<td>nc</td>
<td>nc</td>
<td>calculation of the perturbation growth rate$^2$ nc</td>
<td>application to a numerical forecast model assessment of instability using a single sounding</td>
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<td>Emanuel (1983)</td>
<td>one c-s two WCFR</td>
<td>30–50 km</td>
<td>verified</td>
<td>verified (3 m s$^{-1}$ difference)</td>
<td>nc</td>
<td>nc</td>
<td>calculation of a dry symmetric instability parameter$^3$ nc</td>
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<td>Parsons and Hobbs (1983)</td>
<td>one c-s four rainbands</td>
<td>80 km</td>
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<td>verified (2 m s$^{-1}$ difference)</td>
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<td>verified</td>
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<td>first evidence of roll-like circulation in literature propagation of some band-relative to the mean flow</td>
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<td>Seltzer et al. (1985)</td>
<td>eleven c-s four multi-banded</td>
<td>between 45–115 km (13º deviation)</td>
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<td>not verified</td>
<td>nc</td>
<td>nc</td>
<td>calculation of dry and moist $Ri$ from individual soundings nc</td>
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<td>Sanders and Bosart (1985)</td>
<td>one c-s one snowband</td>
<td>—</td>
<td>nc</td>
<td>nc</td>
<td>verified</td>
<td>nc</td>
<td>study of CSI in presence of forcing</td>
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<td>Wolfsberg et al. (1986)</td>
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<td>nc</td>
<td>verified</td>
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<td>verified</td>
<td>nc</td>
<td>calculation of SPA$^4$ nc</td>
<td>flights along $M$-surfaces</td>
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<td>Emanuel (1988)</td>
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<td>nc</td>
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<td>nc</td>
<td>verified</td>
<td>nc</td>
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<td>Lemaitre and Testud (1988)</td>
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<td>60 km</td>
<td>verified</td>
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<td>verified</td>
<td>verified</td>
<td>verified (from individual soundings)</td>
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<tr>
<td>Source</td>
<td>Number of case-studies (c-s) and band-rays</td>
<td>Spacing between band-rays</td>
<td>Bands parallel to thermal wind flow</td>
<td>Sign of symmetric instability parameters depending on calculation of SCAPET ( \theta_w ) and ( q_w )</td>
<td>General remarks</td>
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<td>Byrd (1989)</td>
<td>27 c-s</td>
<td>98 km (mean value)</td>
<td>verified (2σ deviation)</td>
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<td>nc</td>
<td>calculation of ( q_w ), ( \theta_w ) from individual soundings</td>
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<td>Lemaitre et al. (1969)</td>
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<td>50 km</td>
<td>verified</td>
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<td>nc</td>
<td>calculation of ( q_w ) from individual soundings</td>
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<td>Reiter and Yau (1990)</td>
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<td>50 km</td>
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<td>Lemaitre and Scialom (1992)</td>
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<td>100 km</td>
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<td>verified</td>
<td>use of MANDOP technique to resolve 3D circulations within bands</td>
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**CLOUDBAND DYNAMICAL STRUCTURE**

**TABLE 1.** Continued.

- c-s: case study
- WCFR: wide cold frontal rainband
- nc: not considered or not mentioned in the paper
- SCAPES: standard convective available potential energy (see Emanuel 1983 for details)
- Perturbation growth rate, \( \lambda = -\frac{\partial}{\partial z} \left( \frac{g \theta_w}{\theta_w} \right) \). 
- SPA: slantwise positive area (see Emanuel 1983 and Wobusberg et al. 1996 for definition)
- MANDOP: Multiple Analytical Doppler (see Scialom and Lemaitre 1990 for details)
Figure 2. Surface and 1000-500 hPa thickness map at 00 GMT 13 January 1988.
Lagouvardos et al. (1992). These authors evaluated the frontogenetic forcing, using real data, by solving the diagnostic Sawyer–Eliassen equation, to derive the associated ageostrophic circulations. Their results showed that the frontogenetic forcing had a small horizontal and vertical extension and that, consequently, the direct ageostrophic circulation was confined primarily to lower levels. On the other hand, an indirect ageostrophic circulation was evident at mid-tropospheric levels. This diagnostic analysis also identified the impact and importance of latent-heat release within the narrow cold-frontal rainband, which was found to be responsible for a substantial portion of the observed upright motions. In a similar way, evaporative cooling appeared to play an important role in enhancing the downdraughts located just below the frontal surface.

(c) Radar observations

Figure 4 shows a time series of plan-position-indicator (PPI) reflectivity scans, which were used to investigate the banded structure of precipitation and to follow its evolution with time. In the first radar image, timed at 2143 GMT (i.e. approx. T – 2 h), precipitation is evident in the whole of the range covered by the radar, but is clearly not banded. In the western extremity, the narrow cold-frontal rainband associated with the position of the frontal discontinuity at the surface is just within the limit of the radar’s range. The second picture (Fig. 4(b)) corresponds to the moment of the frontal passage over Brest. The narrow cold-frontal rainband is now evident at Brest, with reflectivity values exceeding 30 dBZ. This narrow band is evident over the whole range covered by the radar and its width is less than 10 km. As observed in earlier studies of narrow cold-frontal rainbands (James and Browning 1979; Hobbs and Biswas 1979), it is fragmented, forming distinct precipitation cores with reflectivity values exceeding 40 dBZ, oriented in a clockwise direction with respect to the front. One of these precipitation cores passed over the sounding station at Brest and gave heavy precipitation during 20 minutes. The
third picture is taken two hours after the frontal passage. It is evident that a region without precipitation characterizes the air underneath the frontal surface, indicating that there was evaporation related to the subsiding branch of the direct ageostrophic circulation. The last PPI image corresponds to 0300 GMT 13 January 1988 (i.e. approx. T + 3.5 h, Fig. 4(d)). The most prominent feature is the precipitating element (C) at a range of about 15 km to the west of the radar. It is oriented at an angle of 20° clockwise from north and its width varies between 15 and 25 km. Element C is located below the
cloudband B which will be studied in more detail in section 6, where also a possible explanation for the formation of the element C will be proposed.

(d) Rawinsonde measurements—frontal description

The very good time resolution of sounding releases from the mobile stations at Brest and Lannion (see Fig. 1) provided an accurate mesoscale data-set which allowed us to construct time–height cross-sections of various dynamical and thermodynamical parameters. Within a time interval of approximately 30 hours, 13 soundings were made from Brest and 14 from Lannion, with a vertical height resolution of 30 m for pressure, temperature and humidity and 150 m for the wind direction and speed. The potential temperature field (Fig. 5(a)) is found to have only weak variations during the frontal passage over Brest; the cross-frontal gradients are not greater than 1 K per 100 km in the vicinity of the frontal discontinuity. Precipitation recorded by a ground station at the sounding site is also shown (Fig. 5(b)). As mentioned above, heavy rain accompanied the frontal passage. Precipitation was also recorded 3.5 h later, corresponding to the passage of the precipitating element C (Fig. 4(d)) over Brest.

The frontal surface is more evident in the along-front wind component cross-section where the wind decreases substantially after the frontal passage (Fig. 6(a)). A narrow
and relatively strong low-level jet is evident just ahead of the frontal surface; its maximum value is 32 m s⁻¹ centred at a height of 1.1 km. A comparison with frontal cross-sections constructed using dropsonde data (see Fig. 4 of Thorpe and Clough 1991) reveals the similarity between the overall structure of these two fields. In particular, the height of the jet maximum and the intense cross-front gradients appear similar. Notice also the small horizontal extension of the jet, which is evident in both figures. Figure 6(b) shows...
the same field, but constructed from the Lannion soundings. The front passed over Lannion two hours later than over Brest (at 0130 GMT 13 January). The remarkable similarity between the two fields indicates the stationarity and two-dimensionality of the system at the mesoscale. However, the low-level jet maximum at Lannion is 3 m s\(^{-1}\) less intense, possibly because of frictional deceleration over land. In both figures, relative humidity contours are superimposed. Note the saturated area ahead of the frontal surface inside the low-level jet and the rapid decrease aloft, both in the warm and the cold sector of the front.

The cross-section of the wet-bulb potential temperature, \(\theta_w\), (Fig. 7) clearly identifies the frontal zone (which contains a well-marked humidity contrast) though the \(\theta_w\) structure is more characteristic of an occluded frontal system than of a classical cold front. This occluded form (V-like shape) where warm air is forced to rise aloft away from the surface could explain the presence of stronger cross-front temperature gradients at a height above 1 km than near the surface (Fig. 5). The \(\theta_w\) isolines are nearly vertical before the frontal passage, in the region where strong line convection was observed. The maximum of \(\theta_w\) at low levels is just ahead of the cold front and within the low-level jet, indicating that this jet provided a source of warm and moist air from the south. This picture is consistent with classical conceptual models of frontal circulation (Browning and Pardoe 1973; Bennett et al. 1989). Hatching denotes areas of convective instability (i.e. \(\partial\theta_w/\partial z < 0\)). The prefrontal boundary layer is convectively unstable; also the whole region underneath the frontal surface in the cold air is unstable.

![BREST SOUNDINGS](image)

Figure 7. Time–height cross-section of the wet-bulb potential temperature, \(\theta_w\), (in 0.5 K intervals). Shading denotes areas of convective instability (\(\partial\theta_w/\partial z < 0\)).

5. **Assessment of Conditional Symmetric Instability**

In this section a comparison is made between observations and the theory of CSI. The following four points will be considered in detail:

1. The relative slope of \(M\)- and \(\theta_w\)-surfaces.
2. The sign of the moist potential vorticity.
(3) The alignment of the cloudband relative to the thermal wind within the unstable region, and the band propagation speed with respect to the mean flow.

(4) The presence of negative correlation between the along-front and vertical-wind perturbations.

(a) Slope of absolute momentum and moist isentropic surfaces

In this subsection, the relative slope of $M$- and $\theta_w$-surfaces is examined. These two parameters allow a study to be made in detail of the frontal structure, thanks to the high time-resolution of soundings performed during this frontal event and the stationarity of the system. The absolute momentum surfaces, $M$, are calculated using the formula $M = v + fx$. In order to define $v$ and $x$, it is necessary to estimate the direction of the thermal wind. Inspection of Fig. 2(b) shows that the thermal wind vector (parallel to the isolines of the 1000–500 hPa layer depth) is nearly parallel to the frontal discontinuity. We therefore measure $v$ as the along-front wind component and $x$ as the distance from the left lateral boundary limit of the domain, perpendicular to the front. The absolute values of $M$ have no significance, since the origin for calculating the distance $x$ has been chosen arbitrarily, and are therefore not shown. Figure 8 shows a cross-section of $\theta_w$ and $M$ isolines. The frontal discontinuity is associated with strong cross-frontal gradients in absolute momentum. In the same region and below a height of 2 km, $\theta_w$-surfaces are nearly vertical, indicating some adjustment to neutrality for upright convection inside the line of strong convection. Regions where the slope of the $M$-surfaces are less than or equal to the $\theta_w$ slope are shaded. We note that these potential regions of conditional symmetric instability are found mainly along the frontal discontinuity, at a height greater than 2 km extending up to 4.5 km. Buckling of the $M$-surfaces (characteristic of active CSI) is not so clear as for the IOP 7 frontal event of FRONTS 87 (documented by Thorpe and Clough 1991 using dropsondes, and by Y. Lemaître and G. Scialom (personal

![BREST SOUNDINGS](image)

Figure 8. Time–height cross-section of the wet-bulb potential temperature, $\theta_w$, (solid lines at 0.5 K intervals) and absolute momentum, $M$, (dotted lines). Regions where $\theta_w$ isolines are steeper than $M$ are shaded.
communication) using the Brest soundings), but this structure will be scrutinized in section 6, where the Doppler radar data will be presented.

The presence of an area of inertial instability is also evident in Fig. 8 for the layer 1–3 km, between 2 and 4 hours before the frontal passage, in the region where the cross-frontal gradient of the $M$-surfaces becomes negative (since $f \partial M/\partial x = F^2$, where $F^2 = f(f + \partial v/\partial x)$). This unstable zone is located on the cold side of the low-level jet (Fig. 6) where the cross-frontal gradient of the $v$-component is strongly negative.

(b) *Wet-bulb potential vorticity, $q_w$*

Calculation of potential vorticity can provide useful information about mesoscale instabilities in the atmosphere. For a two-dimensional inviscid flow, moist potential vorticity defined in Eq. (1) can be written as

$$q_w = F^2 N_w^2 - S^2 S_w^2$$

(3)

where

$$F^2 = f(f + \partial v/\partial x), \quad N_w^2 = \frac{g}{\theta_0} \frac{\partial \theta_w}{\partial z}, \quad S^2 = f \frac{\partial v}{\partial z}, \quad S_w^2 = \frac{g}{\theta_0} \frac{\partial \theta_w}{\partial x},$$

(4)

and air density $\rho$ is taken equal to unity. This is a convenient method of expressing the two-dimensional form of $q_w$, since in this way, we can assess three types of instability related to negative moist potential vorticity, namely,

- Inertial instability ($F^2 = f(f + \partial v/\partial x) < 0$)
- Classical gravitational instability ($N_w^2 < 0$)
- Conditional symmetric instability (both $F^2$ and $N_w^2$ are positive but $q_w$ is negative)

Wet-bulb potential vorticity, $q_w$, was calculated using data provided from the Brest soundings. If the atmosphere is susceptible to convective instability, CSI is not significant because the unstable modes associated with upright convection grow much faster. On the other hand, negative $q_w$ can be associated with inertial instability. So, based on the fact that CSI can persist only in the absence of convective instability, the necessity of distinguishing between various types of instability is evident. Figure 9(a) shows a vertical cross-section of $q_w$ calculated from the two-dimensional formula (Eq. (3)). Shading denotes negative $q_w$. The large regions of negative $q_w$ in the prefrontal and postfrontal boundary layer are due essentially to convective instability (see Fig. 7). Equivalently, the area between 4 and 2 hours before the frontal passage, in the height range between 1 and 3 km is inertially unstable, corresponding to the area of buckled $M$-surfaces (Fig. 8). The areas where inertial or classical gravitational instabilities are present are illustrated with dark shading. Light shading denotes the domain which is stable to both inertial and convective overturning but unstable with respect to CSI. Fragments of negative $q_w$ are evident ahead of the frontal surface, above a height of 3 km. The pockets of negative $q_w$ near the leading edge of the frontal discontinuity are within the area depicted in Fig. 8, where lapse rates measured along $M$-surfaces were greater than the moist adiabatic.

At this stage it would be instructive to compare the $q_w$-field obtained with two different forms of the baroclinic frequency $S^2$. If we assume that the along-front wind component is in thermal-wind balance, then $S^2$ can be written

$$S^2 = \frac{g}{\theta_0} \frac{\partial \theta}{\partial x}.$$

(5)

Moist potential vorticity can then be calculated using this form of $S^2$. Frequently, in an
atmosphere reaching saturation, $S^2$ is taken equal to $S^2_w$, and the criterion for CSI then becomes $F^2N^2 - S^4_w < 0$; but this approximation is not considered here. The result is shown in Fig. 9(b), from which it is evident that now moist potential vorticity is negative in a large region aloft, ahead of the frontal discontinuity. Comparison with Fig. 9(a) indicates that thermal wind balance is not satisfied within this area. Indeed, calculation of the difference between the two terms of the thermal-wind equation in the along-front direction, i.e. $f \partial v / \partial z - (g/\theta_0) \partial \theta / \partial x$, reveals a significant departure from geostrophy over much of the area affected by CSI (Fig. 10). In general, the observed wind is
subgeostrophic, and consequently use of Eq. (5) leads to an overestimate (in magnitude) of negative potential vorticity. Y. Lemaître and G. Scialom (personal communication) have shown how this departure from geostrophy alters the CSI criterion, and that for subgeostrophic configurations $S^2$ should be calculated using Eq. (4) to avoid erroneous overestimations of negative moist potential vorticity. Furthermore, Thorpe and Clough (1991) have shown that regions of negative thermal-wind imbalance (subgeostrophic winds) are associated with anticlockwise CSI rolls. The letter N in Fig. 10 locates such a region of negative imbalance maximum inside the area subject to symmetrical overturnings, ahead of the frontal surface aloft. In section 6, radar data are used to identify an anticlockwise roll within this region.

The estimation of the error made in the calculation of the potential vorticity amounts to a value of $0.2-0.3 \times 10^{-12} \text{s}^{-4}$, which is in good agreement with the error of 0.1–0.2 PVU (since our unit equals $f g \rho / \theta_0$ times PVU) given by Thorpe and Clough (1991) for the calculation of the same parameter with the dropsonde data. This result suggests that the available sounding data can give a rather accurate description of the potential vorticity structure in the frontal region, permitting, especially, the determination of regions where this quantity is negative.

The presence of a CSI region on the leading edge of a frontal surface, but rearward of the surface-frontal location, has been observed (Parsons and Hobbs 1983; Thorpe and Clough 1991; Y. Lemaître and G. Scialom, personal communication), and also predicted by numerical models (Knight and Hobbs 1988). Parsons and Hobbs calculated a dimensionless dry symmetric instability parameter which identified a region unstable to symmetric overturnings at the leading edge of the frontal discontinuity at a height of about 4500 to 5000 m. Thorpe and Clough used dropsonde measurements over the ocean to the west of Brest during the IOP 8 frontal event and found an extended area of negative $q_w$ ahead of the cold-frontal surface aloft. Y. Lemaître and G. Scialom (personal communication) did an analysis identical to ours, but for the IOP 7 frontal event of
FRONTS 87. The results from a calculation of $q_m$, using a network of three sounding stations in Brittany, suggested that there was a region of negative moist potential vorticity near the leading edge of the front, above a height of 3.5 km. On the basis of numerical experiments, Knight and Hobbs proposed a conceptual model for the process that leads to the formation of frontal rainbands (see their Fig. 13). According to their interpretation, a region of negative moist potential vorticity, initially imposed near the surface in the warm sector of the front, is advected aloft along the frontal discontinuity by the ageostrophic circulation. The data provided from FRONTS 87 does not allow us to detect the origin of this unstable air-pocket, but the area of CSI does appear to be located over the ascending branch of the cross-frontal circulation (Lagouvardos et al. 1992).

(c) Wind hodographs

Wind hodographs can be useful in determining whether the observed cloudbands are really parallel to the geostrophic wind shear vector inside the unstable region, and if the bands have a propagation speed equal to that of the mean flow. However, an observed vertical profile in the vicinity of the band might be substantially modified by the roll-like circulation associated with the band itself. A wind hodograph constructed from a rawinsounding released from Brest at 0122 GMT 13 January (i.e. T + 2 h) is displayed in Fig. 11. The radar reflectivity picture at 0146 GMT (Fig. 4(c)) suggests that this sounding has been released in an area of moderate reflectivity, implying that the measured wind has not been seriously affected by the band-related circulation. An Ekman-type spiral is evident in the levels below 1 km, accompanied by a region of strong wind veering, characteristic of the cold air underneath the frontal surface, extending up to a height of

![Wind hodograph](image)

Figure 11. Wind hodograph constructed from the rawinsounding released from Brest at 0122 GMT 13 January 1988. The heights of the measurements are shown in km. The axes parallel to the band are also shown as the direction of the band propagation. Shading denotes the limits of the possible error of evaluation of the band propagation direction from the satellite imagery. The wind intensity and direction at height 3.7 km and the propagation speed of the precipitating element C are also shown.
1.6 km. From this level up to a height of 2.4 km the wind turns in the opposite direction. Between heights of 2.4 and 4.5 km the wind increases in magnitude with height, but its direction remains more or less unchanged, oscillating between 210 and 215 degrees. It seems, therefore, reasonable to consider this flow as the basic flow on which the CSI related flow becomes superimposed.

The vertical wind shear inside the unstable layer (2.4–4.5 km, see previous discussion on the $q_w$ cross-sections) has an orientation of 20–25° clockwise from north, while the band itself is aligned 40° clockwise from north. However, the band seems to be more or less aligned with the vertical shear in the layer between 2.5 and 4 km. This difference of a few degrees has been observed in other studies (see Table 1) and is probably due to the presence of an ageostrophic flow around the frontal surface (Lemaître and Testud 1988; Byrd 1989; Y. Lemaître and G. Scialom, personal communication). Indeed, a well-developed indirect ageostrophic circulation was evident in the mid-tropospheric levels (see Fig. 15 of Lagouvardos et al. 1992), which could result in a clockwise rotation of the cloudband with respect to the thermal wind, consistent with the observations. Recently, three-dimensional CSI numerical investigations by Jones and Thorpe (1992) suggest that the angle between the bands and the thermal-wind direction may be due to viscous effects and that the magnitude of the difference is representative of the importance of the viscous effects. According to Jones and Thorpe’s interpretation, band B of IOP 8 has an anticyclonic tilt.

In addition, CSI theory predicts that bands should have no motion relative to the mean wind. Successive satellite infrared images were used to measure the band displacement, which was found to be approximately 26–28 m s$^{-1}$, from a sector between 220 and 225 degrees. This propagation speed is marked on the hodograph. It is evident that the speed of the band is quite similar to the wind speed at a height of 3.7 km (i.e. nearly half the height of the unstable domain), which suggests that the steering level of band B was at that height, above the cold-frontal surface. Projecting the wind vector at this level and the band propagation vector along the cross-band direction, the propagation vector appears close to the flow speed at half the height of the unstable domain (taking into account errors in the wind measurement and the uncertainty of the band propagation speed evaluated from successive satellite images). This observation seems to be consistent with the propagation of tilted CSI modes described by Jones and Thorpe (1992).

Similarly, the propagation speed of the precipitating element C is shown on the hodograph. Its speed, evaluated from successive PPI images taken every 15 minutes, is 14–16 m s$^{-1}$, from 230 degrees. As shown in section 6, element C extends vertically upwards to a height of about 3 km. Moncrieff and Green (1972) have shown that when the convective available potential energy is maximized, the steady convection moves at the speed of the wind at a steering level which is equal to 0.6 times the depth of the convective region. In the present case, this depth is 3 km, which leads to a steering level at a height of about 1.8 km. The hodograph confirms that the wind at this level is indeed close to the band propagation speed. In addition, the component of the element’s propagation in the cross-frontal direction (the front was aligned 30° clockwise from north) is nearly equal to the frontal propagation speed of 6 m s$^{-1}$, indicating that the precipitating element C was propagating along with the cold-frontal system. This suggests that the process which triggered this convective element is strongly linked to the frontal discontinuity.

\(d\) Correlation between $v'$ and $w'$ fluctuations

A preliminary estimation of the correlation between the $v'$ and $w'$ fluctuations can be made by a single comparison of the time evolution of these two quantities inside the
potentially unstable layer. Since inspection of the along-front component clearly shows that the along-front wind increases with height, a negative correlation between \( v' \) and \( w' \) is expected, consistent with an increase of eddy kinetic energy with time (see Eq. (2)). The vertical cross-sections of moist potential vorticity presented in section 5(b) identify a CSI region in the layer between 4 and 5 km above the ground, at the leading edge of the frontal zone. Since the vertical wind component is not provided directly by the soundings, calculation of \( w \) has been made by integrating the divergence, \( \partial u / \partial x \), from the ground up to height \( z \) equal to 10 km, assuming that the circulation is two-dimensional and that the vertical wind vanishes at the upper and lower boundaries (see also Lagouvardos et al. 1992). The variation with time of the along-front wind, \( v \), and of the vertical velocity, \( w \), at a height of 4.5 km, estimated from the Brest soundings, is shown in Fig. 12. It is evident that band B lies in a region of negative correlation between

![IOP8: z=4500 m](image)

**Figure 12.** The time evolution of the along-front wind component (in m s\(^{-1}\)) and of the vertical velocity (in 10\(^{-2}\) m s\(^{-1}\)) deduced from the Brest soundings, at a height of 4.5 km. The negative correlation scheme between \( v \) and \( w \) is indicated with the plus and minus signs. The position of band B is also shown (rectangular box at the top of the diagram).

\( v' \) and \( w' \) and that strong ascending motions (10 cm s\(^{-1}\)) are also associated with the band. The subsiding motions (\(-5\) cm s\(^{-1}\)) found at the western side of the ascending branch support the idea of a roll-like circulation within the CSI region. Note also, on the one hand, the very good correspondence of along-front wind deceleration (denoted by a minus sign on the \( v \) time variation) with ascending motions and, on the other hand, acceleration (denoted by a plus sign) with subsiding motions. This acceleration-deceleration behaviour of the \( v \)-component within the CSI area is also clearly observed in the vertical cross-section of the along-front wind (A and D in Fig. 6(a)).

A more quantitative description of the component correlation pattern requires a knowledge of the perturbation variables \( u' \), \( v' \), \( w' \), measured relative to the large-scale flow. For this purpose we have assumed that the mean value of the wind at each level of the computational domain is representative of the basic state. Following Bennetts and Ryder (1984) we can then write for the along-front wind component perturbation, \( v' \), the expression

\[
    v' = v \text{ (measured)} - \overline{V^x} \tag{7}
\]
where $\overline{V}^x$ represents the mean of the along-front component on each height level ($V$ in the nomenclature of Eq. (2)). This method is not appropriate at the low levels near the frontal surface where strong cross-frontal wind gradients are present, but is considered to give an accurate estimation of perturbation quantities aloft. The same procedure was used to estimate the two other wind components ($u'$ and $w'$). The resulting data were then used to calculate terms I and II in Eq. (2). As expected, term I was found to be negligible within the area of CSI and term II was positive inside the CSI region (layer 3.1-4.9 km, ahead of the frontal surface). Figure 13(a) represents the mesoscale perturbation field with vectors calculated from the $u'$ and $w'$ data. Superimposed is the streamfunction, $\psi$, calculated by making a vertical integration of the relation $u' = \frac{\partial \psi}{\partial z}$, with the provision that $\psi$ vanishes at the ground. A roll circulation is evident within the CSI region, with the ascending branch situated where band B was formed. Note also that almost the whole roll-like circulation is within the area where term II is positive.

Using the same procedure as before, a composite picture was also obtained using data from the Lannion soundings (Fig. 13(b)); it showed a roll-like circulation at the same height, with the ascending branch observed three to four hours after the frontal passage at Lannion, inside the area of growth of perturbation kinetic energy.

6. THREE-DIMENSIONAL WIND CIRCULATION

(a) Brief review of the MANDOP method

The MANDOP (for Multiple ANalytical DOPpler) retrieval method describes the three-dimensional kinematic characteristics of both stratiform and convective precipitation (Scialom and Lemaitre 1990). Each of the three wind components is expressed as the product of three expansions in the form of an orthonormal set of functions (i.e. one expansion for each spatial coordinate). The radial wind also is expressed in this manner. In order to retrieve the coefficients of the basic functions, the analytical representation of the radial wind field is adjusted variationally towards the observed field subject to constraints imposed by an anelastic continuity equation and a lower kinematic boundary condition satisfied simultaneously by the analytical form of the three wind components.

(b) Results

The wind retrieval was made inside a domain with a length along each side of 100 km and a depth of 6 km, which occupies the major part of the area formed by the overlapping scanning zones of the two Doppler radars deployed during the campaign near Brest (Fig. 1). Six VAD scannings (three from each radar), at 0300 GMT 13 January were used, corresponding to the PPI image of Fig. 4(d), reproduced in Fig. 14. Superimposed is the MANDOP domain (square box) which covers both cloud band B aloft and the precipitating element C at the lower levels. The geometry of the retrieval is chosen so that the y-axis is parallel to the line joining the two radars (oriented 36° clockwise from north). For the advection correction scheme it was supposed that the advecting velocity was equal to the propagation velocity of band B.

The order of the expansion of the orthonormal functions is chosen explicitly to filter out phenomena with a length scale smaller than 15 km (e.g. convective motions). Figure 15 shows the retrieved vectors in a vertical cross-section along the line AD (in Fig. 14) which is almost perpendicular to band B. Two major characteristics are evident in this cross-section:
Figure 13. (a) Time-height cross-section of the perturbation streamfunction, ψ, (solid lines in 5 m²s⁻¹ intervals). Superimposed are the vectors constructed from the u' and w' perturbations; the horizontal and vertical velocity scales are indicated in the upper right-hand corner. (b) As in Fig. 13(a), but for the Lannion soundings.

(1) There is strong reflectivity in the eastern extremity of the domain, related to the rear side of the narrow cold-frontal rainband associated with the frontal passage. Another remarkable feature is the condition of band B in the western side of the domain above the height of 4 km, within the warm sector of the cold-frontal system. Precipitation originating from band B evaporates quickly underneath and consequently the reflectivity
envelope does not extend below 3 km altitude. In addition, this precipitation zone seems to coincide with the ascending branch of the roll, depicted in the sounding data (Fig. 13a,b). Precipitation originating from element C is also evident in the middle of the domain, with a vertical extension up to a height of 3 km. Reflectivity exceeds 20 dBZ within this element.

(2) A roll-like structure is evident in the area of band B, in good agreement with the sounding data. The descending branch of this circulation is not completely retrieved owing to the absence of reflectors (raindrops or snow) near the domain boundary.

According to CSI theory, the ascending branch of the roll has to be parallel to the wet-bulb potential temperature surfaces. Comparison with the $\theta_w$ slope inside the unstable domain (not shown) reveals a more vertical orientation of the ascending branch of the roll. It should be noted however that, for them to have a roll-like circulation, the ascending and descending motions must at some point be directed across the isentropes; which is consistent with the observations.

Figure 16 shows the relationship between the roll circulation and the vertical slope of the $M$-surfaces, constructed from the along-front component provided by the MANDOP method. At low-levels, vectors are aligned nearly parallel to the $M$-surfaces
Figure 15. A vertical cross-section inside the MANDOP domain, along the line AD (see Fig. 14); the horizontal and vertical velocity scales are indicated in the upper right-hand corner. The frontal zone is presented with a dashed line. Also shown are the radar reflectivities in 10 dBZ contour intervals.

Figure 16. Vertical cross-section of the absolute momentum, $M$, (calculated from the MANDOP retrieved along-front wind) along the line AD (see Fig. 14). MANDOP wind vectors are also shown.
while the band-related roll perturbs the \( M \)-surfaces at upper levels, creating a local buckling.

The formation of precipitating element C could be explained on the basis of two different dynamical mechanisms. Figure 13(a) and Fig. 15 show that element C is associated with weak mesoscale ascent, located below the frontal surface and just rearward of the descending region of the direct ageostrophic circulation (Lagouvardos et al. 1992). These slight ascending motions are evident from the perturbation streamfunction field, \( \psi \), estimated from the soundings (Fig. 13(a)), where a kick in \( \psi \) is evident in the layer 500–1000 m, three hours after the frontal passage. One plausible explanation might be the presence of trapped gravity waves between the earth’s surface and the frontal zone, associated with the rapid geostrophic adjustment of the atmosphere. Alternatively, the ascent might be due to low-level convergence in the confluent zone lying between the colder air situated just behind the frontal zone (because of evaporative cooling) and the synoptic-scale airflow in the cold sector. This cold airflow ascends on approaching the cold anomaly region resulting from evaporation, and thus generates the precipitating element C. At the same time, the slightly unstable or neutral layer, located in the descending branch of the CSI roll, could enhance the strength and the vertical extension of convective motions within element C. It is not possible to identify the formative mechanism from an analysis of sounding data alone, because both mechanisms could create a kick in the perturbation streamfunction in the cold air beneath the frontal surface and account for convective elements which are seen to propagate with the same speed as the cold front. However, a detailed study of this frontal case using high-resolution wind-profiler data by Niangoran and Petitdidier (1991) did seem to confirm the presence of trapped gravity waves.

Figure 17(a–c) shows horizontal vector fields derived from MANDOP retrievals. In the low levels (Fig. 17(a)), radars are scanning the cold air below the frontal surface where the airflow has already veered. At a higher level \( z = 4 \) km, Fig. 17(b)) the scanning area is above the frontal surface, where the wind vector is nearly parallel to the front. Note that at both levels the MANDOP retrieved flow is in good agreement with the sounding measurements (barbs).

Figure 17(c) shows retrieved vectors measured relative to the mean flow (26 m s\(^{-1}\), 215°) at a height of 4 km. These can be considered to represent the perturbation field associated with the roll circulation. As suggested in section 5(d) and depicted in Fig. 12, the perturbation field implies an along-front acceleration in the vicinity of the descending branch of the roll and an along-front deceleration in the ascending branch. Therefore, kinetic energy of the bands comes at least partly from the kinetic energy of the large-scale flow, through a negative correlation between \( v \) and \( w \) fluctuations.

7. Conclusion

In this study we have examined the possible role of conditional symmetric instability in the formation of a cold-frontal band observed during the last intensive observational period (IOP 8) of the FRONTS 87 field experiment. The angle between this band and the thermal wind vector inside the unstable layer was approximately 15°, which reflects either the influence of the ageostrophic circulation (Byrd 1989) or the influence of viscous effects (Jones and Thorpe 1992). Several theoretical criteria, as formulated by classical linear theory of CSI have been examined, mainly by using a dense array of sounding measurements taken during this frontal event. The location of an area of negative moist potential vorticity as well as that of an unstable configuration of \( M \) and \( \theta_w \) slopes coincide
Figure 17. (a) Horizontal cross-section inside the MANDOP domain, at $z = 1$ km. The horizontal-velocity scale is indicated in the upper right-hand corner. The wind direction and intensity measured by the soundings are indicated by barbs. (b) As in Fig. 17(a), but at $z = 4$ km. (c) As in Fig. 17(b), but with respect to the mean flow at this level. The plus and minus signs indicate the position of the ascending and the descending branch of the CSI roll, respectively.
Figure 17. Continued.

A detailed analysis of atmospheric motions inside the area enveloping the observed band reveals a well-developed roll circulation, with a descending branch deforming the absolute momentum surfaces, thus creating a buckling region characteristic of active CSI. An observed negative correlation between $v$ and $w$ fluctuations is consistent with the theory of CSI.

Since CSI is often associated with cold fronts, another dynamical mechanism, the frontogenetic forcing, should be examined. An investigation of the interaction between strong forcing and unstable areas with respect to CSI was carried out by Lemaitre and Scialom (1992b) for the IOP 7 of FRONTS 87. This observational evidence of synergy between forcing and symmetric instability is based on recent theoretical and numerical studies by Xu (1989), which emphasize the role of frontogenetic forcing in the number and intensity of cold-frontal rainbands. The fact that, in our study (IOP 8), the horizontal extension of the frontogenetic forcing was small, allows us to conclude that this synergy was of minor importance in this particular case. To confirm this result, it would be necessary to solve the diagnostic Sawyer–Eliassen equation using real data and including the CSI areas, which is a work that deserves further investigation in the future.
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