A study of mesoscale convective bands behind cold fronts.
Part I: Mesoscale organization

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SUMMARY

This paper reports observations made over the North Atlantic of a system of convective cloud bands found within the cold air behind an occluded front. The observations were made from an instrumented aircraft which additionally released a pattern of dropsondes, measuring temperature, relative humidity and wind, as functions of pressure. These data are analysed to establish some of the mesoscale, cloud-scale and microphysical properties of the system.

Here, in part I, the mesoscale structure is described and compared with the predictions of various theoretical models. The convective cloud structure emerges as a consequence of the mesoscale organization but appears to play only a minor role in its creation and maintenance. One of the theoretical models considered, conditional symmetric instability, provides a plausible explanation of the mesoscale observations.

1. INTRODUCTION

This paper describes the results obtained from an investigation into the structure of deep convective, mesoscale cloud bands. The bands were found over the eastern North Atlantic in cold air behind an occluded front and may be classified as one of the several types of approximately two-dimensional features found to be associated with extratropical cyclones, see for example Browning and Mason (1981). Those studied here were orientated parallel to the front and were approximately perpendicular to the 700 mb flow. The bands were 30–40 km wide and had a wavelength of about 80 km. They are to be distinguished from the smaller-scale cloud streets which are orientated almost parallel to the low-level flow and are also commonly associated with unstable air masses.

In general, post-cold-frontal air masses are unstable to moist ascent. The problem, therefore, is not to account for the convection but to understand why the individual convective cells were organized. Two scales of motion are important; cloud scale and mesoscale. If the cloud scale is dominant the bands may be considered as regenerative lines of convection, as for example squall lines. On the other hand, if the mesoscale is dominant, the clouds may be seen as tracers within the dynamical system. Both extremes have been observed and discussed in the literature. Houze et al. (1976) described a relatively narrow (10–20 km), strongly cellular rainband in which a new line of cells formed immediately ahead of the existing line. Matejka et al. (1980) presented observations of a wider (40–50 km) post-cold-frontal rainband which was accompanied by a surface pressure change and a decrease in wet-bulb potential temperature. They concluded that the band was a secondary cold front thereby suggesting that the mesoscale instability was forced by synoptic-scale development.

The present observations have points of similarity with both studies. The bands were cellular (Houze et al.) but had a scale similar to that discussed by Matejka et al. However, a new feature of the present observations is the presence of multiple bands. The organization of the bands appears to result from mesoscale instability within the air mass, rather than the cloud or synoptic-scale forcing implied by the previous two case studies.

2. THE EXPERIMENT

(a) The observing system

The Meteorological Research Flight C130 aircraft is equipped to measure a wide range of meteorological parameters (Nicololls 1978). Of relevance to this study are the Rosemount platinum resistance thermometer for air temperature (to an accuracy of
±0.2°C when observations are made over one-second intervals), the Cambridge automatic hygrometer for ambient dew-point (±0.5°C) and the downward-facing Barnes PRT4 radiometer for sea or cloud top temperature (±0.2°C under optimum conditions but deteriorating rapidly with increasing range). Air motion, detected by a pitot static system and wind vanes on a nose boom, is determined, after the removal of aircraft pitch, roll and yaw, by reference to an inertial navigation system. Horizontal components have an accuracy of ±0.7 m s⁻¹.

The C130 dropsonde facility has been described by Ryder et al. (1983). After ejection, the sonde is suspended beneath a guide surface parasheet, designed for its wind-following characteristics. This provides optimum exposure for the transducers and controls the descent rate at approximately 10 to 12 ms⁻¹, depending on altitude. The sensors are sampled once a second with temperature measured to ±0.5°C, humidity to ±5% in the range 30 to 95%, and pressure to ±2 mb absolute, ±0.2 mb relative, allowing the wet-bulb potential temperature (θₑ) to be inferred to an absolute accuracy of ±0.9°C at 1000 mb and ±0.4°C at 600 mb for 0 ≤ θₑ ≤ 10°C. The sonde is tracked by reference to Loran C navigation aid signals, received at the sonde and retransmitted to the aircraft. Ryder et al. discuss the accuracy of this technique for windfinding and, in the area of this experiment, it was calculated that horizontal winds could be measured to ±0.5 m s⁻¹ when averaged over layers ~600 m deep. Of the eight sondes ejected during the experiment all provided good wind, pressure and relative humidity data but on two of the sondes, marked by asterisks in Fig. 7, the temperature sensors were defective.

The aircraft is equipped with an Ekco E290 3 cm weather radar. The antenna, which has a beam width of 3°, is programmed to scan the 180° sector ahead of the aircraft at one of a number of angles of tilt to the horizontal. This facility, combined with the forward motion of the aircraft, allows a true-dimensional view of the precipitation echo to be constructed. The radar has range compensation and an experimentally determined threshold equivalent to a rainfall rate of about 0.5 to 1 mm h⁻¹ to ranges of 40 km. The radar display was photographed at regular intervals to allow measurements to be made of areas of rainfall exceeding the threshold. Additionally the radar can be set to a range of 200 km, allowing it to be used for control of the experiment. However, in this mode the information was not recorded for subsequent detailed analysis.

Instrumentation for the measurement of microphysical properties is described in part II of this paper.

(b) The synoptic situation

The synoptic situation at 1200 GMT on 19 December 1980 is shown in Fig. 1, based on analyses produced by the Meteorological Office. The vigorous depression situated 500 km west of Scotland was moving slowly east and had begun to fill. The associated fronts were active and brought rain to all parts of the United Kingdom with the heaviest precipitation falling over Ireland, Wales and the southwest Peninsula. Behind the occlusion, in the strong to gale force westerly winds, there was a pool of cold air, marked C in the 1000–500 mb thickness chart, Fig. 1(b). Vigorous convection developed within the cold air, with individual cumulonimbus clouds reaching up to a height of 7 km, their growth possibly being assisted by the large-scale ascent induced by the divergence generated at the left exit of the upper tropospheric jet (Figs. 1(c) and (d)).

Figure 2 is a copy of the infrared satellite picture taken at 0916 GMT, some three hours earlier than the analyses of Fig. 1, and two and a half hours before the start of the observations to be described. On the north, east and south sides of the picture are the trailing edges of the occluded and cold fronts, and in the centre can be seen the vigorous convection associated with the cold pool.

Convection behind a cold front is frequently organized. Often the clouds form open or closed cells typically 50 km in diameter, but sometimes, as in this case, the organization takes the form of mesoscale bands. These can be seen in Fig. 2 between, for example,
Figure 1. The synoptic situation at 1200 GMT on 19 December 1980. (a) Surface pressure (mb); (b) 1000–500 mb thickness (gpdm); (c) 500 mb geopotential height; (d) 250 mb geopotential height. The position of the experiment is indicated by a line.

Figure 2. Satellite photograph (IR) for 0916 GMT on 19 December 1980. (Courtesy of Dundee University)
58°N 15°W and 55°N 13°W, 57°N 14°W and 55°N 14°W, and 57°N 17°W and 55°N 16°W.

From the aircraft observations during the experiment the tops of the frontal, convective and low-level clouds were found to be \( \sim 8500 \) m, \( \sim 7000 \) m and \( \sim 1000 \) m, respectively, the transition between the different cloud types being particularly marked at the rear edge of the frontal zones.

(c) The flight plan

The main objective of the experiment was to study the structure of the convective bands (visible between 12–18°W 55–58°N in Fig. 2) by dropping sondes through them and by making aircraft observations during horizontal traverses. Data from up to five sondes can be received and processed simultaneously on the aircraft. Therefore a sequence of soundings at 20 to 25 km separation can be readily achieved from the aircraft in straight and level flight at 8 km altitude and, by maintaining a true air speed of 150 m s\(^{-1}\), a 300 km traverse can be completed in some 35 minutes. At lower levels the aircraft speed is reduced and at \( \sim 2000 \) m a 300 km traverse takes about 50 minutes. Thus an extensive study involving several horizontal traverses can be completed in a few hours. On this time scale the concept of a constant system velocity can be invoked so that the spatial structure can be derived for many of the important scales. This approach assumes that the system under study is being advected at constant velocity, \( s \), with little change during the observing period. (Some assessment of the validity of this assumption may be obtained from Fig. 3, which shows the development of the system over 22 hours.) The system velocity must be estimated during the experiment if repeated traverses, between points fixed relative to the system, are to be completed. On this occasion, from the movement of the rainfall pattern detected by the airborne radar, and from synoptic-scale data, \( s \) was estimated (in real time) to be 20 m s\(^{-1}\) from 270°T and this value was used to determine the flight pattern. Subsequent more rigorous analysis of satellite data, especially Fig. 3, together with a comparison of the sonde data and the aircraft traverses, Figs. 11 and 12, suggested a value of 18 m s\(^{-1}\) from 270°T.

Figure 3. Satellite photographs (IR) for 2111 GMT 18 December and 0916 and 1903 GMT 19 December 1980. (Courtesy of Dundee University)
Using this revised value of $s$ (18 m s$^{-1}$) the flight pattern was transformed to a 'system-relative' frame of reference, $\mathbf{x} \to \mathbf{x} - s(t - t_0)$, where $\mathbf{x}$ represents the distance from an arbitrary origin on an axis parallel to $s$, in this case eastwards, and $(t - t_0)$ is the elapsed time from an arbitrary time, $t_0$. The result is shown in Fig. 4, where $t_0$ has been chosen as 0916 GMT, the time of the satellite picture. The letters A, B, C, D and E are used here and subsequently to define the cloud bands under study. The presence of upper-level identification somewhat uncertain, these bands is established by other data presented later.

In the evaluation of the measurements it is more convenient to choose $t_0$ nearer to the start of aircraft observations. The result is shown in Fig. 5, where the cloud bands are derived from satellite, sonde and aircraft observations (principally Figs. 2, 7 and 11). Eight sondes were dropped during the first traverse from west to east at ~7200 m above sea level (376 mb). This traverse began at 1150 and ended at 1215 GMT. The second traverse, at ~3500 m above sea level (640 mb),
began at 1235 and ended at 1334 GMT. It was designed to intercept the upper part of the clouds in the bands. A third traverse was at \( \sim 1550 \text{m} \) above sea level (812 mb) during the period 1338 to 1419 GMT. It was the intention that both the latter traverses would provide microphysical data and confirm structural data from the sondes. The positions of these traverses (Fig. 5) were, however, not optimized in system-relative space, the lower traverse ended at \( \sim 250 \text{km} \) rather than continuing to 300 km. This was due to air traffic control constraints.

3. DESCRIPTION OF THE OBSERVATIONS

(a) Sonde and upper-level aircraft data

Typical data, from the sonde \( \delta r \) 180 km in the system-relative frame of reference (Fig. 5), are shown in the form of a temperature diagram and hodograph in Fig. 6. The system velocity

\[ 18 \text{ m s}^{-1} \]

vector (18 m s\(^{-1}\) from 270°T) has been subtracted from the wind at each level and the resultant vector indicates the magnitude and direction of the wind relative to the system, e.g. at 800 mb the wind was blowing towards the north at 2.2 m s\(^{-1}\). For this particular sonde the system propagated northwards relative to the system at all levels, contrary to the definition of system velocity, \( V - s = 0 \) at some level. However, reference to the northward component of wind, Fig. 10, shows this northerly component to be confined to a local region; system velocity refers to the total system.

Most of the region below 600 mb was saturated, the sonde falling through the western edge of band C. Above that height the air became progressively drier. The tropopause was unusually low for the North Atlantic, but was pronounced, and clearly evident at 440 mb. Above the tropopause the air was very dry and at the lower limit of the sonde's operational range. In addition the transducer requires a few tens of seconds (\( \sim 30 \text{mb} \)) to recover from the marked change in environment at ejection (see also Fig. 7).
Figure 7. Vertical cross-section of relative humidity. Dark shading indicates regions with relative humidity >97% with respect to water; medium shading, regions >90% humidity with respect to water and colder than -15°C, possibly indicating ice cloud; light shading, regions saturated with respect to ice. Ice saturation occurs at 85% relative humidity with respect to water at -20°C at 600 mb; at 74% at -30°C at 500 mb. The 60% relative humidity contour is also drawn. At the top of the figure are the cloud top heights inferred from the Barnes PRT4 radiometer, assuming a cloud emissivity of unity.

reinforcing the need for the estimate of dew-point at the top of the profile to be treated with caution; estimates of relative humidity below 10% are at best only accurate to within a factor of two.

The record of the intensity of upward radiation at 8–14 μm provided by the downward-facing Barnes radiometer allows inference of the temperature of the emitting surface. Where, as in this study, the variation of temperature with altitude is known, the height of that surface can be estimated. Emission from intervening water vapour, the finite field of view of the instrument and the presence of ice degrade the accuracy of the estimate especially when the distance between detector and cloud top or sea surface (the emitting surface) is large. Nevertheless, assuming an emissivity of unity, the data shown in the upper part of Fig. 7 provide a useful indication of the altitude of the highest cloud beneath the aircraft during the sonde-dropping traverse. The lower part of Fig. 7 shows the observed relative humidity field. For clarity the figure includes only a representative set of sonde observations. The contours were drawn using all the sonde data and the aircraft data obtained during the lower-level passes. The sonde data were averaged over 10 s, corresponding to a vertical interval of 10–12 mb depending on height. Each set of measurements was made along the trajectory of a sonde and is therefore almost vertical in
this frame of reference and separated horizontally by about 25 km from its neighbours. The downward-facing radiometer data have been used to provide some guidance as to the small-scale structure not resolved by the sondes, particularly where this may be important, as at 170 km. (This feature is also evident in low-level aircraft data, e.g. Fig. 11.)

Indicated humidities above 100% reflect the uncertainty of measurement of this parameter. With this in mind, regions having an indicated relative humidity of greater than 97% are stippled to indicate the probable presence of water cloud. Those areas cooler than -15°C and saturated with respect to ice (>85% at -20°C at 600 mb, >74% at -30°C at 500 mb) are lightly stippled and within these regions the 90% humidity contour is drawn to indicate the probable presence of ice cloud. The radiometer measurements and satellite picture confirm the existence of cloud in some of these regions.

There is evidence that following prolonged exposure in deep cloud the sonde humidity element does not dry out in passing through the base. Where necessary the change in temperature lapse rate from moist to dry adiabatic has been used to establish the lowest cloud base in Fig. 7, and the lower-level relative humidities have been adjusted assuming a well-mixed sub-cloud layer.

The four cloud bands A, B, C and D are evident in Fig. 7 centred at about 80, 150, 200 and 275 km in the system-relative coordinate system (see also Figs. 11 and 12 for aircraft data obtained during traverses). Dry zones, particularly well marked at 700 mb,

Figure 8. Vertical cross-section of the wet-bulb potential temperature (°C). The region below the pecked line is potentially unstable. The data have been meaned over approximately a 25 mb interval and at this resolution have an error of ±0.2 °C. Asterisks indicate sondes with a defective temperature element and for these the error may be slightly greater, see text.
are located at 120 and 240 km. A further cloud-free region, not resolved by the sondes but clearly discernible in the radiometer data (and aircraft data, Fig. 11) is indicated at 170 km. The radiometer data also suggest that bands A and B, and the dry zone at 120 km, are overlaid in part by cirrus cloud.

The field of wet-bulb potential temperature, meaned over 25 mb intervals, is shown in Fig. 8. There is a strong vertical gradient in the vicinity of the tropopause and evidence of a gradual cooling from east to west. As discussed earlier, temperature data are not available from the two sondes marked with asterisks and the average temperature of the adjacent sondes has been used to deduce the $\theta_w$ field at these points. The effect of this approximation is not expected to introduce significant errors in the wet-bulb potential temperatures, and although the absolute values of the minima at 120 and 240 km may be slightly in error, there is no doubt as to their presence, which is primarily due to the humidity minima (Fig. 7). That part of the lower atmosphere below the pecked line in Fig. 8 is potentially unstable.

The velocity components ($u', v'$) are defined relative to the system with respect to axes $(x, y)$, embedded in the system, and moving with the system velocity $s$. The $x$ axis is set parallel to $s$, towards the east and perpendicular to the bands, and the $y$ axis is directed towards the north.

Figure 9 shows the $u'$ component of velocity relative to the system. The strong upper-level flow is the lower part of the westerly jet discernible in Figs. 1(c) and (d). There is also a strong reverse flow in the boundary layer in this frame of reference, similar in

![Diagram](image-url)

**Figure 9.** Vertical cross-section of $u'$. The estimated system velocity, 18 m s$^{-1}$, having been subtracted from the observations. Positive values indicate flow to the east (i.e. right).
many aspects to that expected from surface friction effects. The remainder of the field is fairly uniform and the mean velocity in the central region is less than 1 ms\(^{-1}\) relative to the system velocity, i.e. the speed of the cloud bands. It is therefore concluded that the bands had little motion relative to the air in which they were embedded. It is difficult to be more precise due to the lack of accurate data relating to the movement of the bands, however, the propagation speed was certainly no greater than 2 ms\(^{-1}\) relative to the mean wind. Figure 10 shows the \(v'\) component of velocity relative to the system and the

\[\text{Figure 10. Vertical cross-section of } v'. \text{ Since there was no northerly component of system velocity, the values of } v' \text{ are those observed. Positive values indicate flow to the north (i.e., into the paper).}\]

... contours can be seen to exhibit a pattern with a wavelength not dissimilar to that of the cloud bands, especially the shaded areas at \(\approx 800\) mb. Coherence with cloud bands is poor; this will be examined in more detail in section 4(c) and Fig. 15.

(b) Low-level aircraft data

A summary of the data acquired by the aircraft on the two lower legs is given in Figs. 11 and 12. Figure 11, acquired during the traverse at 3500 m (640 mb), was completed approximately 1\(\frac{1}{2}\) hours after the start of the sonde-dropping leg and the appropriate sonde measurements have been superimposed at their system-relative positions. Because sonde observations were mean measurements in the vertical, whereas aircraft data are displayed as 20 s horizontal averages, it is difficult to make a direct comparison at a given point. Accordingly the sonde measurements are represented by lines covering a range of values, the range being dependent on the fall speed of the sonde, the frequency and
accuracy of the measurements (section 2(a)) and the vertical gradient of the given variable; for example sonde winds are representative of a 600 m vertical interval and in consequence regions of strong vertical shear are not well resolved.

Figure 11(a) shows the relative humidity with respect to water. The minimum at 240 km corresponds to the gap between bands C and D, but other features are less easily related to Fig. 7 because the aircraft flew through the base of the cirrus cloud between 50 and 150 km. However, all the bands A, B, C, D and E are clearly evident in Fig. 11(b), which shows data from the downward-facing radiometer. When the aircraft was in cloud the emitting surfaces were at the same temperature (~−16 °C) as the ambient air, but when the aircraft was in clear air the height of the emitting surface can be inferred by use of a vertical sounding, for example Fig. 6, and this height is displayed on the vertical axis.

Comparisons between the positions of the bands as revealed by Figs. 7 and 11(b) show reasonable agreement; however, there is some discrepancy near 180 km: the separation between bands B and C appears to have increased slightly. On the other hand this may simply be due to non-uniformities in the bands along their length (see Fig. 2). Figures 11(c), (d) and (e) show respectively temperature and $u$ and $v$ components of velocity. Overall there is reasonably good agreement between sonde and aircraft measurements.
confirming that details of the major feature were persistent. The discrepancies that occur are considered to be consistent both with the accuracy of the chosen system velocity and the synoptic and mesoscale development that could occur between the times at which the two sets of observations were made.

Figures 12(a)–(e) give similar data obtained from the last traverse at 1550 m. This was completed approximately 2.5 hours after the sondes were dropped. The temperature and $v$ component of velocity are still in fairly good agreement but the $u$ component indicates that at this level the mean westerly component of wind had decreased by 1 or 2 m s$^{-1}$. The significance of this is not known but, as can be inferred from Fig. 3, the whole system was slowing down.

In Figs. 12(a) and (b) the structure of the bands has been lost. This is because the traverse penetrated the numerous small cumuli that grew between the bands (Fig. 2). The life of these is short compared with the time elapsed from the sonde measurements and therefore considerable changes in humidity at this level would be expected. The radiometer data confirm the presence of numerous small-scale clouds and these obscure the structure of bands B and C although there is some evidence of bands A and D. Additionally the gap between bands C and D is still well marked. As can be seen from Fig. 2 this was the least cloudy of all the inter-band regions.
(c) **Radar data**

Radar data were obtained on the sonde-dropping leg of the flight and typical dimensions of individual precipitation areas were found to be $10 \times 10 \text{ km}$ at cloud base with only a few cells penetrating higher than $4 \text{ km}$. The precipitation in band D was found to be considerably weaker than that in bands A, B and C and, from inspection of the satellite picture (Fig. 2), this might have been anticipated as the band was composed primarily of low cloud. However, rain cells were observed in band E indicating that in the $2^\frac{1}{2}$ hours which elapsed between the time of the satellite picture and the start of the experiment some development, at least in that band, had taken place.

Although it was evident that the rain cells were embedded within the cloud bands, there was no indication of any organization of individual cells. Within the bands, cells appeared to be randomly distributed with all cells having an unicellular structure. No daughter cell growth was observed.

4. **Analysis of data**

(a) **Vertical velocity**

Because the sondes were dropped in a line perpendicular to the cloud bands, the structure in the direction parallel to the bands is unknown and therefore a rigorous derivation of the vertical velocity is not possible. In similar situations, particularly when considering frontal zones, many authors have made the assumption that the synoptic-scale deformation was small ($\partial u_x/\partial x \simeq \partial v_y/\partial y \simeq 0$, where $u_x$ and $v_y$ are the components of the geostrophic wind) and neglected variations along the axis of the front (a rationale for this approach may be deduced from, for example, Hoskins (1978)). Here, although the bands contain embedded convection which might have generated large local deformation fields on the cloud scale, there appears to be little synoptic-scale development (Fig. 1) and the bands are identifiable over a fairly long period (Fig. 3). In consequence the two-dimensional approximation is made and the vertical velocity evaluated from the equation of conservation of wet-bulb potential temperature:

$$\frac{\partial \theta_w}{\partial t} = -u' \frac{\partial \theta_w}{\partial x} - v' \frac{\partial \theta_w}{\partial y} - \omega \frac{\partial \theta_w}{\partial p}$$

(1)

with $u' \frac{\partial \theta_w}{\partial y} \ll u' \frac{\partial \theta_w}{\partial x}$ and $\frac{\partial \theta_w}{\partial t} = 0$. Alternatively, the vertical velocity may be evaluated from the continuity equation

$$\frac{\partial u'}{\partial x} + \frac{\partial v'}{\partial y} + \frac{\partial \omega}{\partial p} = 0,$$

(2)

with $\frac{\partial v'}{\partial y} \ll \frac{\partial u'}{\partial x}$. $u'$ and $v'$ denote departure from the system velocity $s$ (= $18 \cdot 0 \text{ m s}^{-1}$) and the $x$ and $y$ axes are respectively perpendicular and parallel to the bands. The vertical velocity is given by

$$\omega(\theta_w) = -u'(\partial p/\partial x)_{\theta_w}$$

(3)

from Eq. (1) to an accuracy of $\sim 10\%$, or $5 \times 10^{-2} \text{ kg m}^{-1} \text{ s}^{-3}$ whichever is the greater for individual points ($\sim 0.5 \text{ cm s}^{-1}$ in the lower troposphere). The errors in evaluating $u'$ and $(\partial p/\partial x)_{\theta_w}$ are similar. Alternatively

$$\omega(p) = -\int (\partial u'/\partial x) \, dp$$

(4)

from Eq. (2) with $\omega(p) = 0$ at $p = p_0$ where $p_0$ was taken as the level of the tropopause (440 mb), the integral running from $p$ to $p_0$. In consequence the accuracy of $\omega(p)$ deteriorates towards the ground where at worst the cumulative error is $0.3 \text{ kg m}^{-1} \text{ s}^{-3}$ ($\sim 3 \text{ cm s}^{-1}$). Errors due to the neglect of variations along the bands are unknown. The alternative derived values of the vertical velocity are shown in Fig. 13 where, to illustrate the mesoscale perturbation, the mean observed wind in the east–west direction at each level ($u'(p)$) has been subtracted from the data. Velocity vectors are of $(u', \omega)$ where $u' = u - s - \bar{u}'(p)$. 
The velocity vectors in Fig. 13(b), evaluated using Eq. (4), are more coherent than those in Fig. 13(a) evaluated using Eq. (2). In spite of this, comparisons between the two figures are instructive as they give an indication of the validity of the approximations used. For the region to the east of \( x = 125 \text{ km} \) (Fig. 13(b)), \( \int (\partial y'/\partial y) \, dp \), integrated from 977 to 440 mb (= \( w(977\text{mb}) \)), is fairly small, as would be expected close to the surface, indicating that overall the neglect of \( \partial y'/\partial y \) in Eq. (2) was a reasonable approximation; however, for the region west of \( x = 125 \text{ km} \) it is clear that the approximation is less satisfactory. There is an implication of strong divergence parallel to, and at low levels within, band A, although this feature does not seem to be reflected in \( \omega(\theta_u) \) (see Fig. 13(a)). It is possible that the wind field measured by one of the sondes was unrepresentative of
the mesoscale flow, being unduly affected by, for example, one of the many cumulonimbus clouds.

However, overall the two fields (Figs. 13(a), (b)) have a degree of similarity. Close study reveals a weak slanting updraft from 175 km, 750 mb to 140 km, 650 mb and two general regions of descent situated between bands A and B at \( \sim 100 \) km and between bands C and D at \( \sim 225 \) km. Since both these features are found in both figures it is unlikely that they are consequences of the (necessary) two-dimensional assumptions.

Although the fields are rough the main features of the flow pattern are revealed by the superimposed streamlines and two rolls, centred at 125 and 175 km, are evident.

Several different boundary conditions for the streamfunction calculations were tried and although the position and orientation of the rolls did vary by up to 25 km horizontally and 50 mb vertically, we are satisfied that the general behaviour of the solutions shown in Figs. 13(a) and (b) is, although very much smoothed, reasonably representative of the general flow pattern revealed by the velocity vectors in the interior, and that this particular solution is not unduly influenced by the boundary conditions. (The boundary condition in both figures was \( \partial \psi / \partial n = 0 \), where \( n \) is the vector normal to the boundary.)

Study of the velocity vectors at the boundaries of Figs. 13(a) and (b) reveals that there was a net inflow through the sides of the domain. Without variations in the \( y \) direction (into and out of the paper) this implies outflow through the top and bottom (where \( \omega \) would be expected to be zero) and accounts for the open nature of the streamlines at these boundaries. The imposition of \( \omega = 0 \) at these top and bottom boundaries in the calculation of the streamfunction may be considered more realistic, but for these boundary conditions the subsequent large adjustments in the flow dominate the interior solution. Such contradictions are inherent in the two-dimensional approximation that was, in this experiment, necessary. It is hoped to eliminate this problem in future experiments by dropping more than one line of sondes.

In Fig. 14 the mesoscale streamlines (Fig. 13(a)) are superimposed onto the main features of the relative humidity field (Fig. 7). The upward branch of the rolls is centred in the mainly cloudy air in band B and the downward branches are fairly well marked by dry zones at 120 and 240 km. That most of band C is situated within descending air, and some of the dry zone at 120 km within ascending air, may possibly be due to a positional error of approximately 25 km in the streamlines, which is consistent with the uncertainties discussed above. However, the region 850 mb, 200 km is not well represented by the streamfunction (Fig. 13(a)) and some of the individual vectors in this region indicate ascent. The gap in the clouds between what have hitherto been designated bands B and C is difficult to explain. The evidence for separating these two bands was partly from the radiometer data and partly from the interpretation of the satellite picture taken about three hours before the start of the experiment. However, because of the small separation between the bands, it is not possible to resolve the sonde data sufficiently to investigate the wind field in this region. It remains a possibility that the absence of cloud was the result of local variations in the distribution of convective cloud rather than the presence of a distinct gap between separate mesoscale bands of cloud.

For consistency, the discussion in the remainder of the paper will continue to separate bands B and C, but this distinction must be interpreted with caution. The evidence for the separation of bands A and B, and C and D, is much stronger.

(b) Thermal wind

The thermal wind, \( v_T \), may be evaluated from the equation

\[
f \frac{\partial v_T}{\partial p} = -\frac{1}{\rho_0}(\partial \theta_0/\partial x)
\]

where \( f \) is the Coriolis parameter, \( \rho \) air density and \( \theta_0 \) a reference temperature (280 K). The accuracy with which \( v_T \) can be evaluated from Eq. (5) depends almost entirely on the
accuracy of the sonde’s temperature sensor (±0.5 °C). There is a small contribution due to the absolute accuracy of the pressure sensor (±2 mb) but this becomes significant only near the tropopause where the vertical gradient of θ is large and the consequent contribution to the temperature error rises to ±0.2 °C; in most regions the error resulting from uncertainties in the height of the sonde is less than 0.05 °C.

In consequence, over horizontal scales of 25 km, \( \partial v_\theta / \partial p \) may be evaluated to an accuracy of \( ± 2 \times 10^{-4} \text{ m}^2 \text{ s}^{-2} \text{ kg}^{-1} \) rising to \( ± 1 \times 10^{-3} \text{ m}^2 \text{ s}^{-2} \text{ kg}^{-1} \) near the tropopause. Since \( v \) was measured to within 0.5 m s\(^{-1}\) averaged vertically over 600 m, \( \partial(v - v_\theta) / \partial p \) can at best be evaluated to an accuracy of \( 4 \times 10^{-4} \text{ m}^2 \text{ s}^{-2} \text{ kg}^{-1} \), rising to \( 1.4 \times 10^{-3} \text{ m}^2 \text{ s}^{-2} \text{ kg}^{-1} \) at the tropopause. (At the surface \( |\partial v_\theta / \partial p| = 5 \times 10^{-4} \text{ m}^2 \text{ s}^{-2} \text{ kg}^{-1} \) is equivalent to \( |\partial v_\theta / \partial z| \sim 5 \times 10^{-5} \text{ m s}^{-1} \).) Unfortunately, therefore, the accuracy of the observations is insufficient to calculate the departure from thermal wind balance, thereby precluding a direct evaluation of the motion associated with the mesoscale circulations. After comparing the values of \( \partial v_\theta / \partial p \) and \( \partial v / \partial p \) the best that can be said is that, within the accuracy of the data, the atmosphere is nearly in thermal wind balance.

(c) **Flux \( w v_\theta \)**

The inability to calculate departures from thermal wind balance \( (v_\theta) \) is unfortunate as the flux \( w v_\theta \) is an important indicator of mesoscale circulations. Bennetts and Hoskins (1979) noted that in a study of frontal rainbands by Roach and Hardman (1975), \( w \) and \( v_\theta \) were negatively correlated and both Bennetts and Hoskins (1979) and Emanuel (1982)
have shown that $w_{\psi} < 0$ is a necessary condition for the growth of symmetric instabilities. Such instabilities have growth times of typically 3–18 hours.

An alternative estimate of this important quantity is shown in Fig. 15, where the numbers indicate the value of the flux $w\psi''$ where $\psi'' = \psi - \bar{\psi}'(p)$. It is suggested that $\bar{\psi}'(p)$ is an approximation of the thermal wind $v_T (v' = v - v_d)$ and hence that $\psi''$ is an estimate of $v_d$.

In deriving Fig. 15, $w$ has been evaluated from Eq. (3) and therefore each value of $w\psi'' (\times 10^2, \text{m}^2 \text{s}^{-2})$ is an independent estimate made over a 100 mb vertical interval, and is accurate, at 850 mb, to $\pm 10^{-2} \text{m}^2 \text{s}^{-2}$ or 15%, whichever is the greater. The error increases with height as $(\text{density})^{-1}$.

Superimposed on the field of $w\psi''$ are the mesoscale streamlines (Fig. 13(a)) and it can be seen that except for two regions, 100 km, 550 mb and 225 km, 600 mb, $w\psi''$ is negative within the rolls. Measured over the whole domain below 500 mb, $w\psi'' = -0.015 \pm 0.003 \text{m}^2 \text{s}^{-2}$.

5. DISCUSSION

The main characteristics of the observations were:
(i) the bands were orientated north–south;
(ii) the wavelength of the bands was $\sim 80 \text{km}$;
(iii) the bands had no detectable movement relative to the mean flow (i.e. the relative movement was \(<2\text{ m s}^{-1}\));
(iv) the cloud bands were associated with well-defined dynamic rolls superimposed on the mean flow;
(v) ascending and descending regions of air were well defined and inclined at (4 ± 2°) to the horizontal;
(vi) the perturbation velocities, \(\upsilon\) and \(w\), were negatively correlated, implying that the

![Figure 16. Schematic diagram of the mesoscale cloud bands found in the case study. For further details see text.](image)

energy source of the rolls was, at least partly, derived from the kinetic energy of the mean flow;
(vii) convective clouds within the bands were not well organized, and did not appear to produce daughter cells.

Figure 16 is a proposed model of one of the bands. It is based on the observed features described above but some of the finer details reflect conclusions derived in the discussion to follow.

In the model the atmosphere is divided into three layers: a potentially unstable boundary layer; a middle, weak baroclinic zone marginally stable to wet ascent; and an upper stable layer. The middle layer has little vertical wind shear and, relative to it, the upper and lower winds are in opposite directions.

It is suggested that an instability develops within the middle layer and generates a roll perturbation parallel to the horizontal temperature gradient. The ascent is shallow, being inclined at typically a few degrees to the horizontal, and the vertical velocity has a magnitude of a few tens of cm s\(^{-1}\). The rolls do not propagate relative to the mean flow and so the streamlines are parcel trajectories. Consequently there is a substantial vertical movement of air and a band of cloud develops in the region shown.

At the point marked C there is strong low-level convergence and potentially unstable boundary layer air enters the updraught region. On ascent, some of the convective instability is released and cumulonimbus clouds develop, embedded within the cloud band, making the band appear as a loosely linked organization of convective clouds.
Theoretical models providing possible explanations for these observations are:
(i) conditional symmetric instability (Bennetts and Hoskins 1979);
(ii) symmetric CISK in a baroclinic flow (Emanuel 1982).

Two other theories which can give rise to lines of convective clouds, namely wave–CISK (Raymond 1975) and regenerative convection (Ogura and Liou 1980; Miller 1978), can in this instance be dismissed as the slope of the circulation clearly demonstrates that baroclinicity is an important feature of the flow. A fifth theory, cellular convection (Moncrieff 1981), can also be dismissed on the same grounds, but since it contains some interesting ideas on maintaining mesoscale circulations, some aspects of it are briefly discussed.

(a) Conditional symmetric instability (CSI)

The concept of CSI was proposed by Bennetts and Hoskins (1979) to explain the mesoscale rainbands found within frontal zones. Some further work (C. A. Nash, private communication) on the growth of CSI in weak baroclinic zones has suggested that the instability may also develop within the post-cold-frontal region. CSI is a two-dimensional instability which develops as rolls along the thermal wind, i.e. parallel to the thermal thickness (Fig. 1(b)). (The thermal gradient in the experimental area was not well defined but a comparison of Figs. 1(b), 2 and 3 indicates that the orientation of the bands was not dissimilar to that of the thermal thickness lines.) Such rolls are embedded within, and (see e.g. Eq. (14) of Bennetts and Hoskins (1979)) have no propagation relative to, the mean flow.

Application of this theory to the present observational data showed that on scales of 200–300 km the atmosphere was marginally unstable to CSI. Consequently a fuller study was undertaken with a version of the numerical model described in Bennetts and Hoskins. A representation of the thermodynamic fields was used for the initial conditions but, because the \( \theta_v \) field (Fig. 8) already contained roll perturbations similar to CSI, it was unsuitable in its present form. In an attempt to determine the mean properties of the field before the rolls occurred, the mean horizontal and vertical gradients over the domain 60 to 260 km, 850 to 500 mb were calculated (unfortunately there were insufficient data to define the fields from synoptic observations). The gradients were found to be 1·1 K (100 km)\(^{-1}\) and 0·12 K km\(^{-1}\) respectively; however, because of the amplitude of the temperature perturbations, there was considerable uncertainty in these figures and on scales of \( \approx 100 \) km the gradient often exceeded the mean gradient by a factor of two. Consequently a range of values was considered for the numerical studies, 0·75 to 1·5 K (100 km)\(^{-1}\) in the horizontal and 0·1 to 0·2 K km\(^{-1}\) in the vertical. Uniform baroclinic zones having these properties were used as the initial conditions for the model, several different sets of initial conditions in the ranges indicated above being used. The tropopause was simulated by a temperature inversion.

The instability was started with a small velocity perturbation similar to that used by Bennetts and Hoskins. It was about 5 km wide and 500 m high with its centre situated at 800 mb. Initially the growth was exponential and the perturbation expanded until after about eight hours the velocities had attained sufficient magnitude for inertial effects to become important. During this phase, the shape of the instability remained constant and the amplitude of the circulation doubled approximately every two hours but after that time the instability slowly distorted due to the inertial effects, the degree of distortion depending on the precise nature of the boundary conditions and the characteristics of the baroclinic zone.

A typical CSI roll is illustrated in Fig. 17(a). The initial field had wet-bulb potential temperature gradients of 1·5 K (100 km)\(^{-1}\) in the horizontal and 0·1 K km\(^{-1}\) in the vertical and the diagram shows the streamlines after eight hours. The flow is characterized by a narrow region of ascent with broader, weaker regions of descent but the theory provides no natural scale other than that implied by the depth of the domain and the slope of the \( \theta \) and \( \theta_v \) isotherms.

The ascending branch of the roll is inclined to the horizontal at an angle between
that of the $\theta_\circ$ surfaces and that of the absolute vorticity vector. For the range of $\partial \theta_\circ / \partial x$ and $\partial \theta_\circ / \partial z$ considered, this implies an angle of from $2^\circ$ to $5^\circ$ to the horizontal; in Fig. 17(a) the ascending branch is inclined at $2.5^\circ$. The descending branch has a predicted inclination between the absolute vorticity vector and the $\theta$ surfaces and in Fig. 17(a) is inclined at $1.5^\circ$. Further details on this aspect of the structure of CSI are given in Bennetts and Hoskins.

In Fig. 17(b) the streamlines derived from the observations (Fig. 13(a)) are reproduced. It is difficult to assess the inclination of the ascending and descending branches but they are approximately parallel and make an angle of $(4 \pm 2)^\circ$ to the horizontal, similar to that of CSI. However, due to the resolution of the observations and the smoothing necessary to derive the streamfunction, the characteristic narrow region of ascent (Fig. 17(a)) is not apparent in Fig. 17(b).

It is also interesting to compare the magnitudes of the model and observed circulations. A good measure of this is the contour interval shown beneath each figure; the observations have values somewhat higher than the model after eight hours but similar values are obtained in the model after about nine hours integration. Over the range of initial conditions the growth rate of CSI varied considerably and the time taken for the numerical simulation to develop velocities similar to those observed ranged from 9 to 15 hours. If the observed rolls had grown at the same rate, then the initial atmospheric
perturbation from which the bands developed might be expected to have occurred sometime between 2000 h and midnight on 18 December. Comparison between Figs. 3(a) and (b) shows that over this time the bands indeed increased in wavelength, became noticeably larger and deep convection developed.

Before summarizing this section there is one aspect of CSI requiring clarification. As discussed in Bennetts and Hoskins, the energy source of symmetric instability is the kinetic energy of the basic flow. The adjective 'conditional' in CSI indicates that the instability receives additional energy from the release of latent heat, but it is crucial that this release occurs in phase with the vertical velocity and in consequence the release of latent heat takes place over typically 12 hours, the time taken for a parcel to move through the ascending branch of the roll. The convection growing within the band derives its energy from the potential energy of the atmosphere, and releases it over typically 30 minutes, the time taken for a parcel to rise from cloud base to cloud top. There is therefore little direct interaction between the two types of instability. CSI is a feature of the moist baroclinic zone; convection results from the potential instability of the atmosphere. In the observed conditions either instability could have been released independently. However, CSI is capable of organizing the convective release, for as the potentially unstable boundary layer air is drawn into the upward branch of the roll, it will become convectively unstable.

In summary, the theory of CSI predicts that:
(i) the perturbation streamfunction should take the form of symmetric rolls;
(ii) the rolls should have no motion relative to the mean wind;
(iii) the flow is characterized by narrow regions of ascent and broad regions of descent;
(iv) the updraught should be inclined at between 2° and 5° to the horizontal;
(v) the wavelength should be \( \sim 80 \text{ km} \) (mean updraught slope of 3-5° and a domain height of 5 km, see Fig. 13);
(vi) there should be near thermal wind balance (Bennetts and Hoskins 1979);
(vii) there should be a negative correlation between the departure from thermal wind balance \( (u_0) \) and the vertical velocity \( (w) \) (Bennetts and Hoskins 1979).

Each of the above has been examined in the text. Points (i), (ii) and (iv) are well borne out while on (v), (vi) and (vii) there is no serious disagreement with observations. The resolution of the data is insufficient for a detailed assessment of point (iii).

\((b)\) Symmetric CISK in a baroclinic wave

This theory proposes an alternative hypothesis by which latent heat release may be incorporated into mesoscale flows. Using the CISK approach, Emanuel (1982) examined possible 2-dimensional interactions between cumulus heating and internal gravity waves within a baroclinic zone, and found interesting new wave–CISK modes which had maximum growth rates within the mesoscale.

Emanuel suggested that the heating generated by a line of cumulus cloud could be represented by a specified function of the mesoscale vertical velocity at a specified level, i.e. proportional to \( Q(p)\omega(p_0) \) where \( Q(p) \) is the vertical heating distribution function. This is in contrast to CSI where the heating is everywhere a function of the (upward) vertical velocity, i.e. proportional to \( \omega(p) \). In consequence, growing symmetric wave–CISK must propagate relative to the mean flow, a fundamental difference between the two phenomena.

In other respects the two theories have many points of similarity. Symmetric CISK also develops as a roll perturbation but, in contrast to CSI, the mean slopes of the ascending and descending branches are both predicted to be parallel to the absolute vorticity vector. The slopes of the two branches are, however, not uniform, being steeper near the ground than at higher levels. Over the range of baroclinic zones discussed in section 5(a), the predicted mean inclination of both branches varied between 1-5 and 3°, slightly shallower than CSI but still comparable with the observations. In addition the presence of vertical wind shear and background rotation allows the conversion of the
kinetic energy of the basic flow to disturbance energy (Emanuel 1982). In consequence, symmetric CISK in a baroclinic wave exhibits a negative correlation between the perturbation zonal flow and the vertical velocity (as does CSI); in the notation developed in the discussion of CSI, \( w_{v_3} < 0 \).

It is, however, important to note that symmetric CISK is a two-dimensional instability which results from the interaction of lines of cumulus clouds and internal gravity waves. While, on the mesoscale, details of the along-band distribution of the clouds are unimportant, in practice the cumulus will occur as distinct cells. Parcels of air approaching the band from the side will either enter a cloud and rise vertically within it, or pass between clouds and remain at the same level, or possibly even descend slightly. The two-dimensional mesoscale flow described by Emanuel is therefore an along-band average of two radically different flows.

In the interpretation of sonde data one of the inherent assumptions is that a series of sondes dropped on a line perpendicular to the band can reveal the major features of the mesoscale flow. For CSI this assumption is valid provided the sondes do not fall through a convective cloud. Symmetric CISK, however, exhibits significant along-band variations and therefore for this instability a series of sonde measurements cannot, except fortuitously, represent the along-band average necessary to reveal the structure fully.

In summary, in conditions such as were observed on 19 December 1980 the theory of symmetric CISK in a baroclinic wave predicts that:

(i) the mean perturbation streamfunction should take a similar form to that of a symmetric roll;
(ii) the rolls should propagate towards the warm air (Emanuel 1982);
(iii) the updraught and downdraught should be inclined at between 1.5 and 3° to the horizontal;
(iv) the wavelength should be \( \sim 130 \text{ km} \) (mean updraught slope of 2.25° and domain height of 5 km, see Fig. 13);
(v) there should be near thermal wind balance;
(vi) there should be a negative correlation between the perturbation velocity \( (v_3) \) and the vertical velocity \( (w) \).

Apart from a slightly longer wavelength, (iii) and (iv), many of the points are similar to those discussed in the section on CSI and the conclusions are equally applicable here. The main difference is in the prediction of a propagation velocity. To determine this would, in general, require a measure of the heating function driving the convection. This was not measured, but fortunately the wavelength of the observed bands fell within the parametric region where the phase speed is insensitive to this quantity, and, given the observational measurements, Emanuel's theory predicts a propagation velocity of \( \sim 6 \text{ m s}^{-1} \) relative to the mean flow (18 m s\(^{-1}\)), directed towards the west. The observations, although not conclusively proving that the bands had no motion relative to the mean flow, nevertheless indicate that it was small, less than 2 m s\(^{-1}\).

(c) The cellular convective model

A theory of steady convection organized into mesoscale cells was advanced by Moncrieff (1981) and Fig. 18 outlines the essential elements of the theory. The circulation is divided into two regions. Region 1 is convectively driven while region 2 is forced by the kinetic energy generated within the first, the necessary thermodynamic equilibrium between a parcel and its environment being maintained by radiation, evaporation and surface fluxes of heat and moisture. The model is highly simplified and requires two assumptions: that the basic atmospheric state is the same in both regions; and that ascent and descent are reversible. The theory, which is evaluated by comparison with the clockwise circulation centred at 200 km, 650 mb (Fig. 14), may be divided into three parts. The first relates to the general structure of the convective cell; the second to the prerequisite that a parcel of air moving round the circulation outlined in Fig. 18 be in thermodynamic
equilibrium with its environment; the third to whether sufficient kinetic energy is generated within region 1 to drive the circulation.

The first and third are not examined for the reasons outlined earlier, but it is convenient to begin discussion of the thermodynamic cycle at the point marked X in Fig. 18 (taken to be the point 180 km, 600 mb in the observational dataset). As the cloud was advedted away from region 1 (towards the east) the mean temperature of the cloud was influenced by both radiation loss and solar absorption. The latter was difficult to quantify but fortunately, in the present study, was negligible (<10 W m\(^{-2}\) at noon) as the experiment took place at 56°N a few days before winter solstice.

To determine the infrared radiation emitted from the cloud, the numerical model described in Roach and Slingo (1979) was used. This showed that the net radiational loss was 86 W m\(^{-2}\), the value being almost independent of detailed microphysical properties of the cloud, provided that it was optically thick. Observations indicated that the upper cloud was, in fact, optically thick and therefore the main source of uncertainty lay in translating the radiation energy loss into a cooling rate. For a cloud layer 50 mb thick, e.g. at 225 km, 600 mb, the mean air temperature would fall by 0.17 K h\(^{-1}\), but there is considerable uncertainty as to the actual thickness of the cloud layer and it would be unreasonable to expect this figure to represent the observed conditions to better than a factor of two.

To follow the subsequent history of the parcel of air requires a knowledge of the environment through which it passed. This is not known, as the evolution of the bands was not measured. However, a typical sounding in the vicinity of the right-hand loop is given in Fig. 6(a) and is sufficiently representative thermodynamically, but not in cloud structure, to discuss the evolution of the cloudy layer as it cooled and sank. In consequence, its path is more likely to follow the path XY rather than to move horizontally and then fall steeply, as indicated by Moncrieff. Since parcels of air in this region contain typically 1, and at most 1.5 g m\(^{-3}\), equivalent liquid water content (see part II), they could descend to, respectively, ~750 and ~800 mb. Application of parcel theory to Fig. 6 indicated that a radiation heat loss sufficient to reduce the wet-bulb potential temperature by 2 K was required to induce descent to 800 mb. At a cooling rate of 0.17 K h\(^{-1}\), this suggests a time of ~12 h to complete the descent. With a mean horizontal velocity at cloud top of ~1 m s\(^{-1}\) relative to the bands, this implied a horizontal scale for the roll of ~50 km.

Further descent (beyond Y) was not possible for, in the absence of condensed water substance, the parcel would conserve potential temperature and become warmer than the environment, thereby destroying the necessary thermodynamic balance. Consequently the
return branch of the circulation must be located at a height of about 800 mb. Some evidence of this can be found in Fig. 8 where a tongue of low $\theta_w$ air is situated between 800 and 900 mb, 200 and 250 km.

The last part of the thermodynamic cycle required returning air to be modified so that by the time it re-entered region 1 it was again convectively unstable. Moncrieff suggested that this could be achieved by surface fluxes of heat and moisture, but since the return path was well above the boundary layer, this mechanism would seem unsuitable here. An alternative mechanism is, however, available. Figure 8 shows that the advection of the tongue of low $\theta_w$ air at the bottom of the circulation created low-level potential instability (for example the region below 850 mb at 175 km). This could cause overturning of the boundary layer and returning air, and subsequently the boundary layer air, would reach 800 mb and take part in the convection in region 1.

In summary, an examination of the properties of a parcel of air within the mesoscale circulation shows that descending air is probably nearly in thermodynamic equilibrium. For the theory of cellular convection this is crucial as the driving force, kinetic energy generated by cumulonimbus convection, is weak and unable to force a circulation not in thermodynamic equilibrium. For the other two theories previously discussed, CSI and baroclinic–CISK, thermodynamic equilibrium is not a necessary condition: in both there is sufficient energy released in the ascending branch of the circulation to overcome the atmosphere's stability to descent (see e.g. Bennett and Hoskins (1979)). In consequence, the presence of radiative cooling, evaporation and surface fluxes of heat and moisture, which effectively combine to reduce the atmosphere's stability to descent, provides an important mechanism which can assist many forms of mesoscale circulation. In particular, such a mechanism could assist the development of both CSI and baroclinic–CISK.

6. CONCLUSIONS

This paper reports an experimental flight which investigated a system of mesoscale convective bands situated a few hundred kilometres behind an occluded front. The data obtained have been compared with two theories: conditional symmetric instability (CSI, Bennett and Hoskins 1979); and symmetric CISK in a baroclinic wave (Emanuel 1982). After a detailed investigation, CSI is considered to give the best account of the observed characteristics, the predicted propagation speed and longer wavelength of symmetric CISK making that theory less likely to have been applicable.

Certain aspects of cellular convective theory (Moncrieff 1981) were also examined as the idea that radiative cooling, evaporation and surface fluxes of heat and moisture could combine to reduce the atmosphere's stability to descent is attractive. Since the individual mechanisms are not dependent on the precise dynamics of the circulation, they could assist many forms of mesoscale circulation, and on this occasion may have enhanced the growth of CSI.

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