The structure of rainbands within a mid-latitude depression

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SUMMARY

A case study is presented of a winter depression over the British Isles in which extensive banded structure was observed within precipitation ahead of the surface warm front. Measurements of the mesoscale airflow and precipitation structure of the rainbands were made using a variety of radar techniques together with multiple radiosonde and aircraft observations. The measurements were made over the sea to avoid the confusing effects of topography. The dominant rainbands were oriented parallel to the surface cold front and were typically 100 km wide. They moved with a velocity faster than the underlying warm front. For the most part the bands were characterized by clusters of weak small-scale convective cells due to the release of potential instability produced where tongues of relatively dry air of low $\theta_w$ in the middle troposphere overran low-level moist air undergoing slantwise ascent above the warm frontal zone. Although there was the usual large-scale, and thermally-direct, circulation associated with the active warm front, the air which ascended as small-scale convection within the rainbands entered a region of weak cold frontal baroclinicity, whereupon it participated in a thermally direct circulation of its own. This led to each rainband having a rearward-sloping anvil cloud canopy characterized by ascending air with colder drier air descending beneath. Precipitation falling from the canopy evaporated within the underlying drier air thereby probably intensifying the descending branch of the circulation. Very large ageostrophic winds were measured in association with these circulations.

The important ingredient responsible for the convective nature of the rainbands appears to have been the incursion of tongues of relatively dry air of low $\theta_w$ in the middle troposphere above the moist warm-sector air in a region where the resulting instability could be realized by large-scale ascent. Although the potential instability was very weak in the present case, the origin of the rainbands appears to have been similar to that of pre-frontal squall lines. The intensity of the convection within rainbands depends on the stability but the very existence of any precipitation in the first place depends on other dynamical factors leading to widespread ascent.

1. INTRODUCTION

In recent publications considerable evidence has been presented for the existence of sub-synoptic scale rainbands in association with frontal systems and in the warm sectors of mid-latitude depressions. Browning and Harrold (1969), in a study of a wave depression over the British Isles, showed that in addition to uniform precipitation ahead of the warm front, bands of heavy rain occurred both ahead of the surface warm front and within the warm sector. Some bands (Type B1) were approximately parallel to the warm front: they occurred ahead of the surface position of the front. Other bands (Type B2) were oriented at a large angle to the warm front: these were observed mainly within the warm sector but also extended some distance ahead of the surface warm front. It was argued that the B1 bands were composed of clusters of small-scale convective cells associated with the release of potential instability in the warm air by the large-scale frontal ascent, while the B2 bands were the result of a similar release of potential instability, enhanced by features of
the topography (with which the bands appeared to be strongly associated). Further studies by Harrold (1972) have confirmed that both types of bands are common features of the precipitation pattern in mid-latitude frontal systems over the British Isles. In studies of occlusions over the United States Pacific coast and over New England, Elliott and Hovind (1964) and Kretzberg and Brown (1970), respectively, demonstrated the occurrence of similar rainbands, and suggested that their orientations were related to the vertical wind shear in the region of the convection. Type B1 and B2 orientations were present in Kretzberg and Brown’s study in the absence of obvious orographic effects and in the absence of a surface warm sector. In a study over the sub-tropical ocean off Japan, Nozumi and Arakawa (1968) showed rainbands to be a common feature in the warm sectors of young depressions. These bands, which had a B2 type orientation and occurred in the absence of topographic influences, were considered to be similar to pre-frontal squall lines observed in the United States (e.g., Newton 1950).

The conventional model of a frontal depression, as first formulated by Bjerknes (1919), does not describe these mesoscale aspects of the precipitation distribution. As a first step towards formulating a more realistic model, Harrold (1972) has analysed the large-scale flow in the warm air in a way that provides a useful framework in which to consider the detailed structure of the rainbands. Differential advection within the warm air results in the generation of regions of potential instability, while the large-scale dynamical ascent leads to the release of the instability. The main features of Harrold’s model are shown in Fig. 1. This shows a warm sector depression at a mature stage. The production

![Diagram](image_url)

Figure 1. Model portraying the generation of potential instability within a mid-latitude depression (redrawn from Harrold 1972). Horizontal and vertical sections are shown in (a) and (b), respectively, the section in (b) being along the line AB in (a). The stippled arrow in each diagram represents the flow relative to the system within the conveyor belt. The hatched arrow represents the flow of a layer of potentially colder air in the middle troposphere. Isopleths in (b) are of wet-bulb potential temperature (θ_e) labelled in °C. The principal regions of potential instability (∂θ_ε/∂z < 0) are generated near the surface cold front and also where the hatched arrow overruns the stippled arrow.
of baroclinic precipitation occurs mainly within a tongue of moist warm air; the moist flow is derived from a region of small-scale convective mixing (within the planetary boundary layer) in the warm sectors and sometimes from as far away as sub-tropical anticyclones. This air accelerates ahead of the cold front as a well-defined flow typically 200 km wide and 2 km deep before ascending above the main warm frontal zone. (See the broad stippled arrow in Fig. 1.) Green, Ludlam and McIlveen (1966) and Palmén and Newton (1969) have described this flow as being the most significant ascending flow in systems of large-scale slope-convection. In studies of the general circulation it has been compared to a conveyor belt transporting large quantities of heat and westerly momentum polewards and upwards, and the label ‘conveyor belt’ will henceforth be used to describe it in this paper.

The top of the conveyor belt is normally defined by the base of mid-tropospheric air warmed and dried by earlier descent. Convection within the conveyor belt as it travels over a relatively warm sea surface results in its having a higher wet-bulb potential temperature than the base of the overlying dry air. As these flows ascend in the large-scale baroclinic field the potential instability is realized, leading to their mixing and to the eventual removal of the upward decrease of wet-bulb potential temperature. Thus in general there is a decrease in small-scale convection as the flows ascend to the forward side of the warm front. Fig. 1(b), showing a section along the line AB in Fig. 1(a), demonstrates this process. Harrold has suggested that flows of this type can be identified in the majority of mid-latitude baroclinic disturbances. The configuration of the conveyor belt depends on the synoptic scale motion field and so the precipitation distribution depends upon the interaction between the large-scale dynamic ascent within the conveyor belt and the potential instability at its upper boundary. The purpose of the present paper is to demonstrate this interaction by means of a detailed case study of organized rainbands.

2. The observations

The observations were made on 18 January 1971. They were obtained for the most part over the sea in order to minimize ambiguities introduced by topographical disturbance of the airflow similar to those encountered by Browning and Harrold (1969). The observations were centred on the Isles of Scilly (hereafter referred to as S), a small group of low-lying islands 40 km west of the mainland of south-west England.

The principal measurements of precipitation patterns were made using a Plessey 43S radar the characteristics of which are summarized in the footnote*. The radar was located at S. Autographic raingauges at sites on S allowed accurate calibration of echo intensity in terms of rainfall rate to be made. The radar was programmed to operate automatically in a sequence of RHI and PPI modes, to provide the horizontal distribution of precipitation out to 200 km range and the detailed three-dimensional structure out to 50 km. The broad (2°) beamwidth of the radar limited quantitative precipitation estimates to within about 50 km. At longer ranges useful semi-quantitative data on the pattern of precipitation were obtained. Autographic raingauges on the mainland enabled the precipitation structure to be studied in regions where orographic influences were observed.

Measurements of the airflow pattern associated with the rainbands were obtained using the three-dimensional mesoscale wind-finding technique described by Hardman, James and Goldsmith (1972). Radar reflectors were dropped from two aircraft: an Argosy from RAF Benson and the Varsity of the Meteorological Research Flight. These enabled the wind field to be determined over a volume 4 km deep × 80 km wide × 4 hours long (equivalent to a distance of 300 km relative to the system). The spatial resolution was about 30 km. The 4-hour observational period corresponded to the entire period of surface precipitation and the dropping line was centred about 30 km upstream of S. Fig. 2(a)

* Characteristics of Plessey 43S radar: wavelength 10 cm, peak transmitted power 650 kW, pulse repetition frequency 600 or 275 s⁻¹, pulse lengths 0.15 or 1.5 μs, aerial gain 38 dB, half-power beamwidth nominally 2° in vertical and horizontal, minimum detectable signal 108 dBm.
Figure 2. (a) Plan distribution of data sources on 18 January 1971. (b) Time-height cross-section showing the nature and distribution of data on 18 January 1971. The data from Camborne and from the aircraft have been displaced to the position of the Isles of Scilly according to the rainband velocity.
FRONTAL RAINBANDS

shows the plan distribution of soundings. When displaced relative to the moving system the soundings from the Argosy/Varity aircraft are distributed fairly evenly through the system (see area EFGH in Fig. 3). The automatically opening radar reflectors dropped by the Varsity were modified to carry a temperature sensor and transmitter allowing simultaneous wind and temperature profiles to be obtained. Two precision tracking radars and radiosonde telemetry were located at S.

A vertical section of the mesoscale properties of the wind field was also provided in the vicinity of S during the period of precipitation using a ground-based X-Band (3 cm) pulsed Doppler radar in the conical scanning mode. This radar, described by Bahns and Whyman (1966), was operated in a manner similar to that used in other recent studies of frontal precipitation (e.g. Browning and Harrold 1969), with further refinements for optimizing accuracy as outlined by Browning (1971).

Temperature, humidity and wind measurements were made using conventional radiosondes released sequentially at Camborne (03808) Aberporth (03502) and Larkhill (03743) at intervals of approximately 1.5 to 2 hours. Temperature and humidity measurements were also obtained from ascents released hourly at S in the periods before and after the surface precipitation. Temperature, wind and turbulence data in the vicinity of S were also obtained with the Canberra aircraft of the Meteorological Research Flight using Doppler radar and an Inertial Platform (Axford 1968).

The locations of all these data are shown in the four-dimensional framework provided by Fig. 2(a) and (b).

3. THE HORIZONTAL STRUCTURE OF THE RAINBANDS

Extensive rainband structure was observed during the passage of a frontal system on 18 January 1971. The synoptic situation at 1200 GMT is depicted in Fig. 3. This shows a

Figure 3. Surface and 1,000–500 mb thickness analysis for 1200 GMT, 18 January 1971. Stippled shading represents the satellite cloud distribution. S represents the position of the Isles of Scilly. The rectangle ABCD corresponds to the area of Fig. 4(b). The area EFGH corresponds to the rather smaller region covered by dropsonde data from the Argosy/Varity aircraft (see Fig. 2(a)).
partly occluded frontal system, the point of occlusion passing just north of S. The position of the surface warm front in the vicinity of S coincides with a check in both the rate of fall of surface pressure and the rate of rise of the dew-point; the position of the surface cold front coincides with the advent of a pressure rise and wind veer (refer ahead to surface observations plotted in Fig. 7(b)). The superimposed cloud distribution was derived from an ESSA 9 satellite cloud photograph at 1055 GMT.

Fig. 4(a) shows individual 0° elevation PPI displays at 0749, 0910, 1037 and 1314 GMT during the passage of the system. Fig. 4(b) is a composite diagram derived from PPI data at 20 min intervals, individual photographs having been superimposed using the observed velocity of movement of precipitation features. Three main rainbands can be identified with an orientation of about 200°–020°. The labels: Rainbands I, II and III, and Uniform Precipitation, on the PPI displays in Fig. 4(a) are reproduced in locations corresponding exactly to their position in the composite diagram – Fig. 4(b). The lack of precise correspondence between the labels and the echo bands at long range is due mainly to the combined effects of beam geometry and earth curvature. At 200 km range the base of the beam, assuming a 0° horizon, would have been at a height of 2-3 km, whilst with a 1° horizon the base would have been at 5-8 km.

It is evident from Fig. 4(a) that the bands contained considerable small-scale cellular structure. This is shown most clearly by the quantitative data obtained with the 43-S radar at short ranges (Fig. 4(c)). The precipitation pattern was most cellular in Rainbands II and III. Towards the leading edge of the echo, where the rainbands lost their identity, the radar echo was more uniform. The forward edge of this more uniform precipitation was oriented along 170°–350°, approximately parallel to the surface warm front (Fig. 3). This orientation, which was a consistent feature of the leading part of the echo pattern and surface rainfall distribution, is the same as that of the upper cloud edge (Fig. 3).

The width of each of the three main rainbands was of the order of 100 km. This structure was reflected in the mesoscale wind field, particularly above the main warm frontal zone. Fig. 5, derived from the dropsonde data 30 km upwind of S, shows the positions and orientations of axes of maximum veer and backing in the horizontal wind field at 200 m intervals between 3 and 4 km. Maximum veer is observed to correspond closely to the gaps between the bands, whilst axes of maximum backing correspond to the centres of Bands II and III. Band I is not well represented in Fig. 5, its passage occurring just before the start of the dropsonde pattern; however, the Doppler radar winds showed that maximum backing was associated with Band I also.

An x-t diagram has been constructed to demonstrate the movement of the rainbands in relation to the frontal structure (Fig. 6). The orientation of features in this diagram represents their velocity. According to Fig. 6, the leading edge of the uniform precipitation advanced steadily at the same speed as the surface warm front (16 m s⁻¹) while the rainbands moved at a higher speed (22 m s⁻¹), more nearly equal to that of the surface cold front. Thus elements in the rainbands developed close to the location of the surface warm front, moved ahead of it, and dissipated at the leading edge of the precipitation, the entire process taking about 10 hours. The cellular nature of the precipitation within the rainbands is shown in Section 4 to be the result of small-scale convection. The trend for the precipitation to become less cellular as the rainbands overtook the warm front (Fig. 4(c)) is shown to be consistent with all of the potential instability being realized during the large-scale slantwise ascent above the warm front (Fig. 1). The subsequent tendency for the mesoscale areas of heavier precipitation, which had originated as clusters of convective cells, to persist while degenerating into areas of more uniform precipitation is consistent with their remaining within a region of large-scale ascent.

Rainfall intensity within the bands was mainly 1-4 mm h⁻¹, whilst between the bands it fell to below 0-5 mm h⁻¹. Over the mainland, Fig. 6 shows that orographic effects become prominent, e.g., at x = 80 and 170 km downwind from S. These locations correspond to the high ground of the Penwith Peninsula and Dartmoor. Rainfall totals for the entire system varied between 2-5 and 11 mm at mainland sites compared with 5-5 mm at S.
Figure 4(a). PPI displays from the 43S radar at S, on long (200 km) range and 0° elevation, showing the extent of precipitation echo at full gain at 0749, 0910, 1037 and 1314 GMT.

Figure 4(b). Schematic distribution of precipitation within the area ABCD in Fig. 3, as inferred from the qualitative long-range PPI data from the 43S radar. The extent of surface rain is shown stippled. The further extent of precipitation aloft is lightly stippled. The leading and rear edges of the dense high-level cirrus canopy is indicated by hatched shading. The time scale indicates when the corresponding parts of the pattern passed over S.
Figure 4(c). Detailed distribution of surface rainfall within part of the area covered by Fig. 4(b), as inferred from the quantitative short (50 km) range data from the 43S radar. The hatched, cross-hatched and solid shading represents three levels of radar reflectivity corresponding approximately to rainfall intensities of 0·2, 0·8 and 3 mm hr$^{-1}$, respectively. Thick lines indicate the main features of the precipitation pattern as shown in Fig. 4(b). The time scale indicates when the corresponding parts of the pattern passed over S.

Figure 5. Axes of maximum veering and backing of the wind at levels 3·0, 3·2 ... 4·0 km, as derived from the aircraft dropsonde data within the area EFGH in Fig. 3.

4. THE VERTICAL STRUCTURE OF THE RAINBANDS

The structure of the rainbands changed little during the time taken for each band to pass over a fixed location. This suggests that a two-dimensional steady-state model may be used, at least as a first approximation, to describe the vertical structure of the rainbands. This was achieved by deriving time-height sections through the band structure. On a longer time scale, as the rainbands moved through the entire system, they did evolve, but this does not invalidate the use of time-height sections. Where possible, data along the
bands were averaged in the parallel-band direction to smooth out detailed features. The set of diagrams comprising Fig. 7 shows different aspects of the vertical structure derived from the data sources displayed in Fig. 2.

Fig. 7(a) is a time-height section of the precipitation echo intensity and relative humidity. The surface rainfall from an autographic open scale raingauge at S is shown at the bottom of the Figure. Echo associated with Bands I and II extended to 9 km; Band III echo extended to only about 4 km. At the leading edge of the system, Band I merged with an area of more uniform precipitation, the downward extent of which was limited by total evaporation within a large region of very dry air. The extensive precipitation aloft was succeeded after 0930 GMT by zones of precipitation separated by dry tongues. One of these tongues (at 6 km) extended beneath an anvil-like canopy protruding rearwards in association with Band II. The canopy was identified visually by means of aircraft observations.
Figure 7(a). Time-height cross-section showing the intensity of the precipitation echo and the distribution of relative humidity during the passage of the frontal system over S. The echo pattern has been averaged along the rainbands over a distance of ±30 km either side of S. The hatched shading represents four levels of radar reflectivity roughly equivalent to the rainfall rates shown in the key. The dashed lines denote 25 and 75 per cent relative humidity with respect to ice. Regions of less than 75 per cent relative humidity are stippled. The positions of the surface cold front (SCF) and surface warm front (SWF) are indicated. Surface rainfall rate averaged over 10 minute periods, as measured by an autographic raingauge on S, is plotted at the base of the diagram.

Figure 7(b). Time-height cross-section showing wet-bulb potential temperature, $\theta_w$, at 1°C intervals. Regions of potential instability ($\partial \theta_w/\partial z < 0$) are stippled. Surface observations at S are plotted along the bottom of the Figure. The extent of cirrus as given by aircraft and satellite observations is indicated at the top.

Figure 7(c). Time-height cross-section showing $u_w$, the component of the wind transverse to the warm front, relative to the warm front, at intervals of 5 m s$^{-1}$. Regions where the wind component is directed rearwards relative to the warm front are stippled. Regions where the wind component exceeds the velocity of the front by more than 10 m s$^{-1}$ are hatched.
FRONTAL RAINBANDS

Figure 7(d). Time-height cross-section showing \( u_v \), the component of the wind transverse to the rainbands, relative to the bands, at intervals of 5 m s\(^{-1}\). Regions where the relative humidity is less than 75 per cent are stippled.

Figure 7(e). Time-height cross-section showing the distribution of vertical air motion. Solid lines represent vertical velocity, at 10 cm s\(^{-1}\) intervals, as derived accurately from the dropsonde data and the Doppler radar conical scans. Dashed lines represent vertical velocity, as inferred approximately from Fig. 7(c) and (d) (see text). Regions of ascent are stippled. The small rectangular frames represent regions of vertical convection with local updrafts of order 1 m s\(^{-1}\), as detected by the Doppler radar scanning vertically.

and by means of the 43S radar, the associated radar echo being depicted in Fig. 7(a) with its rearward tip at 1130 GMT. The rear edge of this canopy was also identified on the satellite cloud pattern (Fig. 3) where it had a similar orientation to the rainbands. Further very dry air was encountered behind Rainband III; the increase in relative humidity above it at a height of 4 km accompanied a shallow cloud canopy associated with Band III. This canopy resembled that associated with Band II but the ascent within it was presumably too weak and too shallow to produce precipitation particles large enough to be detected by radar.

Fig. 7(b) shows the distribution of wet-bulb potential temperature (\( \theta_w \)). The main synoptic features, as indicated by strong vertical gradients of \( \theta_w \), are an intense warm frontal zone with a slope of about 1:200 reaching the surface at about 1300 GMT and a cold frontal zone rising from 2 km at 1230 GMT to 5 km at 1500 GMT.

Dry air limiting the forward side of the precipitation area (Fig. 7(a)) was located in and below the main warm frontal zone (Sawyer 1955). Similarly very dry air at low levels behind Band III extended in and below the main cold frontal zone. Fig. 7(b) shows that the air just beneath the cold frontal zone had a \( \theta_w \) as low as 7°C. The sea surface temperature within 500 km upwind of \( \delta \) was between 11 and 13°C and this gave rise to a rather shallow layer of air of relatively high \( \theta_w \) close to the sea surface. Although the resulting strong potential instability in this region extended to a depth of about 2 km (stippled shading in Fig. 7(b)), this instability was not being realized owing to subsidence and the dryness of the overlying air of low \( \theta_w \). The passage of the surface cold front, defined by a pressure kick and wind veer, occurred at 15 GMT about 2 hours behind the leading edge of the cold air at 2 km. Further potential instability was present in the cold air below the main warm front but again dry subsiding air restricted convection to a shallow layer.
In the warm air the main feature of the thermal structure, as seen in Fig. 7(b) and later in Fig. 9(a), is a rather deep tongue of relatively cold air with low $\theta_w$, centred at 1230 GMT, within which there was little vertical gradient of $\theta_w$. Probably this tongue was elongated parallel to the system of rainbands. Henceforth we shall refer to it as the Cold Tongue with Capital letters, to distinguish it from other tongue-like features. Weak potential instability characterized large parts of the Cold Tongue, especially near its forward edge, beneath the Band II canopy (Fig. 7(a)). The top of the Cold Tongue was marked on its rear side by a warm baroclinic zone sloping down from 8–10 km at 1230 GMT to 6–8 km at 1400 GMT where it merged with the main cold frontal zone. On its forward side a weak cold baroclinic zone bounded the top of the Cold Tongue, sloping up from 5–6 km at 1030 GMT to 8–10 km at 1230 GMT; despite its weak appearance, it will be shown later that this cold baroclinic zone was associated with a well-developed transverse circulation.

One of the primary aims of this study was to describe the three-dimensional airflow patterns associated with the precipitation structure. To this end, detailed patterns of vertical air motion have been derived using the dropsonde and Doppler radar techniques described in Section 2. Limitations imposed on both these techniques restricted measurements to below a height of 4 km on this occasion. In order to provide a satisfactory description of the airflow relative to the rainbands it has been necessary to use less direct methods to derive vertical motions above 4 km. As mentioned before, the configuration of the bands observed by radar has suggested some degree of two-dimensionality. Two methods are possible using this assumption. If one neglected gradients along the bands one could obtain horizontal divergence from $D = \frac{\partial u}{\partial x}$, where $u$ is the velocity component along the x-direction, perpendicular to the band orientation. Integration of the continuity equation would then allow the vertical velocity to be calculated. This method becomes unreliable at high levels because of the compounding of error when integrating with height. We have adopted an alternative method using surfaces of constant wet-bulb potential temperature. This again relies on the two-dimensional assumption but errors are not cumulative in the vertical. A difficulty is, however, that relative horizontal flow must be used to derive vertical velocities and this requires a knowledge of the motion of the $\theta_w$ surfaces. The thermal features above the main warm frontal zone are assumed to have had the same orientation, and to have moved with the same velocity (22 m s$^{-1}$), as the rainbands; the warm frontal zone and features beneath it are assumed to have moved at the slower velocity (16 m s$^{-1}$) as derived in Section 3.

Fig. 7(c) shows the vertical section of $u_\theta$, the wind component perpendicular (and relative) to the main warm front, and Fig. 7(d) shows $u$, the wind component perpendicular (and relative) to the bands. Fig. 7(c) shows a well-developed transverse circulation associated with the warm front, the downslope component coinciding with the tongue of dry air beneath the base of the front (Fig. 7(a)), and the ascending branch giving a maximum value of $u_\theta$ at the top of the warm frontal zone. Fig. 7(d) shows that above the warm frontal zone the flow was predominantly overtaking the bands from the west, particularly in the (upslope) region above the upper warm frontal zone to the rear of the Cold Tongue and the (downslope) region behind the main cold front. However, significant rearward relative flow was also present, particularly within the Band II canopy where the rearward flow exceeded 15 m s$^{-1}$. A similar but much weaker flow was present above the main cold front within the cloud canopy associated with Band III. The air within the Cold Tongue was stagnant relative to the bands except for some forward motion beneath the Band II canopy and weak rearward motion beneath the upper warm frontal zone. Double arrows in Fig. 7(d), indicating the axes of these flows, are supported by the distribution of relative humidity (see stippled shading). A comparison of Fig. 7(b) and (d) suggests that each baroclinic zone was associated with a thermally direct circulation; (Fig. 7(b) shows $\theta_w$ rather than $\theta$ and so strictly one should refer ahead to Fig. 9(a) for the distribution of baroclinicity).

As the flow in the Band II canopy ascended above the forward edge of the Cold Tongue
its direction backed, its velocity increased, and it developed an increasingly strong component into the plane of Fig. 7. The cloud associated with this canopy was limited to the forward edge of the Cold Tongue. Possibly the canopy flow retained a rearward component and descended on the rear side of the Cold Tongue; more likely, however, the air at the tip of the canopy disappeared out of the section in association with a horizontal deformation field with its axis of dilatation parallel to a trough line along the summit of the Cold Tongue.

Vertical velocity, as derived from Fig. 7(c) and (d), is shown in Fig. 7(e) by the dashed isopleths. Where the axes of the transverse relative flow, as indicated by the double arrows in Fig. 7(c) and (d), departed from the $\theta_w$-surfaces of Fig. 7(b), this has been taken as an indication of the inadequacy of the two-dimensional, steady state, assumption and the flow has been inferred from the double arrows rather than from the $\theta_w$-surfaces. The solid isopleths in Fig. 7(e) correspond to the more quantitative dropsonde vertical velocities averaged over 80 km along the rainbands and supplemented by Doppler radar measurements. A horizontally extensive region of significant ascent existed over a depth of 2 km near the top of the main warm frontal zone, lying along almost its entire length. This may be identified as the conveyor belt flow. Ascent was not uniform, but increased in association with the rainbands, reaching 20 cm s$^{-1}$ in Band II where the strong mesoscale ascent extended upwards into the Band II canopy. Below the main warm frontal zone weak descent predominated until 1100 GMT. In the region of evaporation at the base of the main precipitation overhang the descent, as inferred from Fig. 7(c), exceeded 10 cm s$^{-1}$.

Two major regions of intermittent small-scale convection (with updraughts of order 1 m s$^{-1}$ over distances of order 1 km) were detected by the vertically pointing Doppler radar in the regions indicated by rectangular frames in Fig. 7(e). One of these regions, between altitudes of 4 and 6.5 km, was associated with the deep precipitation of Band II. The other region, between 1.5 and 3.5 km, contained pockets of shallow convection mainly associated with Band III. Confirmation of convection was provided by the cellular nature of the echo tops on the 43S radar RHI display and by the Canberra aircraft at 5.5 km using the inertial platform. Moderate turbulence and cumulus-scale updraughts up to 5 m s$^{-1}$ were encountered by the Canberra in Band II and, although no reliable mesoscale vertical velocities were measured by the aircraft, data from the other sources described suggest that the small-scale convection in Band II was embedded within a region of mesoscale ascent of the order of 20 cm s$^{-1}$. Air which ascended as small-scale convection continued to ascend within the canopies where it constituted the ascending branches of the transverse circulations described earlier. Where the Band II canopy flow ascended over the forward edge of the Cold Tongue, precipitation fell from the canopy, evaporating in and cooling the descending drier air beneath. This process enhanced the descent, and transverse-band component, of the air within the Cold Tongue above the conveyor belt to provide a mechanism for the maintenance of the potential instability and its realization within Band II.

Band III was associated with the main cold front which had a feeble ana-cold frontal circulation with moist air ascending weakly just above the front and dry air descending beneath the front. Fig. 7(e) shows that the main mesoscale ascent associated with Band III occurred close to the nose of the cold front at 1230 GMT where small-scale convection was observed by the Doppler radar. Presumably this convection was fed by potentially unstable air breaking through the cold frontal zone as the nose of the cold front began to rise above the warm front. Actually, the convection box in Fig. 7(e) covers a large area ahead of this nose, extending into the rear part of Band II; however, evidence from the variability of dropsonde fallspeed suggests that strong small-scale convection was most extensive towards the rear of Band III where visual observations also revealed cumuliform cloud.

Important features of the vertical structure have been combined schematically to give the model in Fig. 8. The conveyor belt is shown as the large arrow ascending above the main warm frontal zone. The leading portion of this flow produced a fairly uniform canopy
of precipitation at the leading edge of the system which, because of evaporation within subsiding dry air beneath, failed to reach the surface over the first 100 or so km. Rainbands I, II and III were characterized by small-scale convection where air in the conveyor belt undercut a tongue of dry cold air situated at middle levels in the warm sector. Bands II and III contained clusters of active convective cells at the time they passed over the Isles of Scilly; however, the convection within Band I had decayed by the time it passed overhead. Rainband II, and to a much smaller extent the other bands too, were accompanied by rearward canopies of stratiform cloud, fed by the small-scale convection from the underlying conveyor belt, and associated with slantwise ascent above drier descending air. The transverse circulation associated with each canopy was thermally direct.

5. Ageostrophic flow within the rainbands

Normally it is difficult to determine the magnitude of ageostrophic winds on the mesoscale; however, the high density of upper air measurements in this study enables us to achieve this, at least partially. The time records of surface pressure and temperature aloft at (effectively) a single station have been transformed to a spatial cross-section assuming that the patterns were advecting over the station in a steady state, and this has enabled the component \( u_g \) of the geostrophic wind normal to the movement \( (260^\circ) \) of the pattern to be derived. Fig. 9(a) shows the time-height distribution of dry-bulb potential temperature.
0, at S, and the resulting distribution of the geostrophic wind component towards 350°. Isotachs are shown at 20 m s⁻¹ intervals; errors due to inaccuracies of individual radiosondes are likely to be about ±10 m s⁻¹. The Cold Tongue within the warm air mass is shown to be the major feature, with a change in the geostrophic wind component across the associated upper trough of as much as 60 m s⁻¹ within 150 km (∼2 hr on the time scale of Fig. 9).

The actual, measured, wind component (v) towards 350°, plotted in Fig. 9(b), reproduces a similar trend in velocities to that in Fig. 9(a), with a distinct maximum near the tip of the Band II canopy coinciding with the geostrophic maximum. Although some of the wind soundings from which Fig. 9(b) is derived may be somewhat unrepresentative owing to the ascent within the rainbands being concentrated within kilometre-scale updraught cells, the component of the actual wind is seen to differ from geostrophic by an appreciable margin. Indeed, it is often as little as half the geostrophic wind, e.g., 20 compared with 40 m s⁻¹ at the tip of the Band II canopy (position C₂ in Fig. 9(b)). Assuming a steady state pattern of flow and neglecting friction, the observed maximum ageostrophic wind component of 20 m s⁻¹ would imply large inertial terms, given by (u ∂u/∂x + v ∂u/∂y) where the x axis is in the direction of the pattern movement. Wind measurements at positions C₁ and C₂ in the Band II Canopy (Fig. 9(b)) show that ∂u/∂x ≈ 0. No measurements of ∂u/∂y are available but, with v ≈ 20 m s⁻¹, a value of 10⁻⁴ s⁻¹ for ∂u/∂y would account totally for the observed ageostrophic component.

6. Discussion

A case study has been presented of a winter-time depression over the British Isles in which extensive banded structure was observed within the precipitation above the warm frontal zone. The rainbands showed a considerable resemblance to banded precipitation systems observed elsewhere (see Introduction). The dominant bands were oriented parallel to the surface cold front, were typically 100 km wide, and they moved with a velocity faster than the main warm frontal zone above which they were situated. The bands were characterized by small-scale convection due to the release of some of the potential instability produced where a tongue of relatively dry air of low θ_w at middle levels overran the low-level moist air in the warm sector of the depression. For lack of a better term this middle-level air has been referred to as the Cold Tongue. The low-level moist flow, referred to as the Conveyor Belt, ascended as a fairly shallow flow above the warm frontal zone and, as it did so, the potential instability at its upper boundary became exhausted and the small-scale convection ceased. At the same time the rainbands lost their identity and, with the continuing large-scale ascent towards the leading edge of the precipitation system, they evolved into areas of relatively uniform precipitation elongated parallel to the warm frontal zone.

The important ingredient responsible for the convective nature of the rainbands appears to have been the incursion of the low-θ_w air above the conveyor belt in a region where the resulting potential instability could be realized by large-scale ascent, in this case above the warm frontal zone. It is not clear whether the organization of the convection into bands was the result of pre-existing banded structure within the potentially cold air aloft or to other (unknown) dynamical factors leading to a banded pattern of large-scale ascent within the underlying conveyor belt flow. Of the two explanations, that involving pre-existing bandedness within the pattern of potential instability is favoured by the finding that the bands lost their identity when the potential instability became exhausted. Alternatively it is possible that the existence and spacing of the bands is some function of the cumulonimbus convection itself rather than the result of streakiness previously introduced into the pattern of either potential instability or large-scale ascent. If this is the case then it would be necessary to think of the band development as leading to the disturbances in the fields of wind and baroclinicity rather than vice versa. The origin of the rainbands is probably similar to that of pre-frontal squall lines, except that the latter are produced in
circumstances where the potential instability is more pronounced. One of the rainbands in this study resembles, on a rather smaller scale, the winter-time pre-(cold)-frontal precipitation bands in the USA described by Omoto (1965). As stressed by Omoto, although the intensity of the convection within the bands depends on the stability, the existence of any precipitation depends on other dynamical factors leading to widespread ascent.

An interesting aspect of this study is the existence of separate mesoscale transverse circulations in association with each of the rainbands. Although there was the usual large-scale, and thermally direct, circulation in the vicinity of the active warm frontal zone, the air which ascended from the top of the conveyor belt as small-scale convection then entered a region of weak cold frontal baroclinicity, whereupon this air participated in a thermally direct circulation of its own. This led to each rainband having a rearward-sloping anvil cloud canopy characterised by moist ascending air with colder, drier, air descending beneath. As might be expected the entire circulation aloft associated with each of the rainbands closely resembles that within squall lines (Newton 1950) and ana-cold fronts (Browning and Harrold 1970). The coexistence of several rainbands gave rise to what Kreitzberg and Brown (1970) have called multiple mesoscale hyper-baroclinic zones.

The best-developed anvil canopy was found bounding the forward edge of the Cold Tongue in association with Band II. Its transverse circulation was far more strongly developed than that of Band III which was associated with the main cold frontal zone. Possibly the ana-cold frontal ascent associated with Band III was inhibited to some extent by descent within the overlying Cold Tongue. The exact significance of such circulations is not clear but it is reasonable to expect that they might intensify and prolong the life of the convective rainbands by evaporative chilling of the low-$\theta_w$ air within the Cold Tongue and by locally enhancing the differential advection so as to regenerate the potential instability.

The above study has been made on the basis of data obtained over the sea with the specific aim of avoiding the confusing effects of topography. Over land the organization of rainbands may be disrupted and, in certain circumstances, topographically-triggered small-scale convection may lead to the establishment of fresh convective rainbands downwind of major hills (Yanagisawa 1961; Browning and Harrold 1969). The influence of topography can be expected to be greatest for rainbands within the warm sector. The rainbands in the present study were situated mainly ahead of the surface warm front and they may have been protected to some extent from the effects of topography by the layer of high stability and wind veer within the warm frontal zone. The susceptibility of convective rainbands in various circumstances to orographic effects is an important research task for the future.

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