Observations of the mesoscale sub-structure in the cold air of a developing frontal cyclone


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SUMMARY

Observations from 58 dropwindsondes released in a mesoscale array during the FRONTS 92 experiment are interpreted in the context of satellite imagery to derive the mesoscale structure and evolution of parts of a frontal cyclone developing over the eastern North Atlantic. A conceptual model involving the intertwining of ‘dry intrusion’ and ‘cloud head’ flows is corroborated and is used to provide the framework for interpreting the detailed mesoscale behaviour. In the cold air, two distinct dry-intrusions were responsible for two cold fronts, trailing south-westwards from the tip of the cloud head. Both were surface features at the beginning of our study but the leading one evolved into an upper front with mid-level convection as the dry intrusion responsible for it overran the warm sector. Areas of both high and low potential vorticity were indicated within the dry intrusions. Upon encountering a critical level of zero system-relative velocity at the top of the moist boundary layer, the dry-intrusions’ arrival was associated with the development of multiple dry and moist laminar near the top of the boundary layer. The vertical wavelength of the laminar was about 1 km and they extended over 200 km in the front-normal direction, with a slope of typically 1 in 60. Although most parts of the laminar were subaturated, their circulations combined with the double structure of the dry intrusions to produce multiple shallow cloud-lines within the boundary layer. These formed as an extension of the south-westerly tip of the cloud head associated with the developing cyclone. Possible mechanisms for generating the observed structures are discussed.

KEYWORDS: Cloud head Dry intrusions Frontal cyclone Laminar Mesoscale structures

1. INTRODUCTION

Browning and Roberts (1994) proposed a conceptual model of a developing frontal cyclone or wave, which integrates a number of well-known features such as the warm conveyor-belt (Harrold 1973), cold conveyor-belt (Carlson 1980), split front (Browning and Monk 1982), frontal fracture (Shapiro and Keyser 1990), dry intrusion (Young et al. 1987) and cloud head (Bottger et al. 1975). With the omission of the warm conveyor-belt, their model reduces to that shown in Fig. 1, in which the cyclone develops in the region where two diffuent hammer-head-shaped flows interlock. One of these flows (stippled), corresponding to Carlson’s cold conveyor-belt, consists mainly of moist ascending air which leads to the characteristic cloud-head with well-defined outer edge as seen in satellite imagery. Part of the stippled flow that extends behind the low centre gives rise to the bent-back front at Stage II of Shapiro and Keyser’s frontal-fracture process. The other flow (not stippled) consists of dry-intrusion air that has descended from the upper troposphere and/or lower stratosphere. The leading edge of this dry intrusion corresponds to a cold-frontonal zone. The two flows mirror each other in developing rather well-defined convex leading edges.

The overall structure implied by Fig. 1 sets the scene within which a number of mesoscale features occur. At first sight the mesoscale structure within a developing cyclone may appear to be rather disorganized. However, satellite and radar imagery, interpreted in the light of a conceptual model and output from a numerical weather prediction (NWP) model, can often be used to identify a coherent mesoscale organization. The purpose of this

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Figure 1. Highly idealized conceptual model showing the intertwining around the cyclone centre (L) of two rather symmetrical flows: the dry intrusion and the cloud-head flow. The cloud-head flow consists of cold-conveyor-belt air and air from the base of the warm conveyor-belt (the warm-conveyor-belt flow, associated with the polar-front cloud-band, is not shown here). The flows are drawn relative to the cyclone system.

The main contribution to this paper comes from a set of 58 dropwindsondes released from a research aircraft in the vicinity of a developing cyclone over the eastern North Atlantic during the Fronts 92 project. These are analyzed together with satellite imagery to reveal the fine structure within the dry intrusion, or rather two dry intrusions in close succession. The analyses reveal in detail how the dry intrusions interacted not only with the moist air beneath and ahead of them but also with the moist air that was recirculating cyclonically to the north of the cyclone centre, within the south-western tip of the cloud head. Relationships are established that are believed to have some generality for a class of developing frontal cyclones. The cyclone in the present study deepened fairly rapidly over a short period but it was not an unusually intense development. For 18 hours the cyclone retained an almost constant central pressure while the cloud head began to form (Fig. 2).
Then, between 12 UTC 27 April 1992 and 00 UTC 28 April 1992, the cyclone deepened by 12 mb and, by the end of this period, the cloud head had acquired a characteristically well-defined convex outer edge. The dropwindsondes were released between 12 and 21 UTC 27 April; shortly afterwards the system was observed by land-based radiosondes and radars.

2. The Dropwindsonde Data

The flight track of the Meteorological Research Flight C-130 aircraft during Intensive Observational Period 3 of FRONTS 92 is shown in Fig. 3, with positions and times of soundings marked. The dropwindsondes (hereafter referred to simply as dropsondes or sondes) were deployed from a height of 6.1 km. The sondes benefited from an improved pressure-sensor but were otherwise similar to those used in the FRONTS 87 project (Thorpe and Clough 1991). Details of observations made in the FRONTS 92 project are presented in Hewson (1993). In this context, winds were evaluated using 30-second averaged cross-chain LORAN winds. The accuracy of windfinding is dependent upon the LORAN signal quality, which varies with location, atmospheric conditions and transmission effects. From signal monitoring, the accuracy was estimated to be about 1 m s$^{-1}$ throughout most of the area, as in FRONTS 87. The consistency between adjacent soundings indicated an accuracy at least as good and perhaps even better than this in most places. However, it deteriorated to between 2 and 3 m s$^{-1}$ occasionally (i.e. in parts of some soundings) in the extreme west of the area. Below cloud which has a relatively high water content, as in the lower parts of Figs. 20(a) and (b), some lag is evident in the response of the humidity sensor; in addition, measured air temperatures may be too low because of wet-bulb effects. These deficiencies should be noted, but they are not important to most of the features discussed here, which occurred at middle levels in subsaturated air.

Data analyses were prepared for the dropsonde observing-area by assigning a standard analysis time of 18 UTC 27 April, the main synoptic hour nearest to the mid-time of the observations. This allowed a corresponding surface-pressure analysis to be derived from available synoptic observations. The system-relative locations of the dropsondes at 18 UTC, as shown in Fig. 3, were calculated by displacing actual sounding positions with a system velocity of 19 m s$^{-1}$ on a heading of 073° from true north. This velocity was our early estimate of the motion of the low centre, which turned out to be 3 m s$^{-1}$ less than
the mean velocity of the cloud head in the satellite imagery. This uncertainty in the system velocity will affect the details of subsequent analyses but not the main conclusions.

Observations from 14 ships and 18 land stations were used to carry out a surface-pressure analysis employing a two-dimensional smoothing-spline algorithm (see Thiebaux and Pedder 1987) in order to provide maximum differentiability. Isentropic analyses were also derived from three-dimensional (3-D) objective analyses of the drops soundings. The analyses were based on the successive-correction algorithms of Pedder (1993), which are computationally more economical for large 3-D data-sets, and for the specification of filtering properties. For these analyses a linear-trend basic state was taken for scalar variables (temperature, wind components and relative humidity) and the method of successive corrections applied to the deviations.

3. MESOSCALE STRUCTURE DURING RAPIDLY DEEPENING PHASE

(a) Overview of surface frontal pattern in relation to cloud head

The surface pressure and frontal analysis for 18 UTC 27 April is shown in Fig. 4, with simplified cloud-analysis superimposed. The pressure analysis has been carried out objectively as outlined in section 2 above. The frontal analysis uses the detailed evidence presented in later sub-sections. The cloud analysis is based on the Meteosat imagery. The edge of high cloud associated with the polar-front cloud band is identified by a cusped line; so, too, is the north-western edge of the cloud head, part of which extends out of the diagram to the north-east. We do not discuss the polar-front cloud band in this paper, as it was largely above and ahead of the main region of development; rather, we concentrate on the structure and airflow associated with the cloud head and with the two cold fronts that
are shown extending south-westward from it. These fronts were better defined in terms of wet-bulb potential temperature ($\theta_w$) than dry-bulb potential temperatures ($\theta$); in some ways it would be better to refer to them as cold $\theta_w$-fronts but for simplicity we shall call them cold fronts. As we shall show, the cold fronts and the cloud head were closely related; the top of the cloud in the cloud head lowered towards its south-western tip where it merged with the shallow cloud bands associated with the cold fronts. The observed lowering of the cloud head towards its south-western tip resembles the schematic illustration in Fig. 1 of Carlson (1980).

(b) Visualization of the broad 3-D pattern of airflow

Key features of the airflow pattern associated with the cyclone are represented in Fig. 5 by relative flow in isentropic surfaces as derived from the dropsondes. This diagram is intended only to introduce the reader to the overall flow pattern; further isentropic analyses are given in Fig. 9. The three surfaces shown in Fig. 5 are:

(i) $\theta_w = 12^\circ\text{C}$ moist isentropic surface. This shows a saturated flow corresponding to part of the warm conveyor-belt (WCB), 200 km wide, ascending from low levels (<1 km) in the warm sector to above 4 km over the warm-frontal zone ahead of the cyclone. (The upper part of the WCB, which was responsible for the polar-front cloud band (see Fig. 4) is not depicted here.)

(ii) $\theta_w = 8^\circ\text{C}$ moist isentropic surface. This is one amongst several surfaces in the range $6 < \theta_w < 10^\circ\text{C}$ which represent the mainly saturated cold-conveyor-belt (CCB) flow. Relative to the system, air in the CCB travelled north-westwards, within and beneath the warm-frontal zone. The CCB then fanned out, the northern part turning towards the north and then north-east, and the southern part turning south-westwards within the cloud head just to the north of the cyclone centre. Initially this flow ascended up the isentropic surface but the ascent decreased as it turned south-westwards parallel to the slope of the isentropic surface. To the extent that the pattern could be considered to be in a steady state, the flow would be inferred to have then descended down the isentropic surface near the tip of the cloud head. In reality, in a rapidly deepening cyclone, a steady state is a poor assumption: the isentropic surfaces were being deformed with time so as greatly to diminish, and possibly to eliminate, the implied descent in this region.

(iii) $\theta = 25^\circ\text{C}$ dry isentropic surface. This shows the north-westerly relative flow characteristic of the deep layer of dry air descending behind the cyclone. Part of it overtakes the cloud head, and part overran Cold Front 2 (CF2) and Cold Front 1 (CF1), and turned to become an ascending south-westerly flow. Initially the flow was very dry (relative humidity <30%) but the relative humidity (RH) increased as the initially dry air overtook the system. Further evidence of this relatively dry flow protruding ahead of CF1 is shown in Figs. 10 and 14.

(c) The two cold fronts CF1 & CF2

The double cold-frontal analysis in Figs. 4 and 5 is consistent with both satellite imagery and information gained from the dropsondes. In particular, the double structure is evident in the 950 mb $\theta_w$-pattern shown in Fig. 6. It is also evident in the almost north-south vertical section along Run 5 shown in Fig. 17. Double structure in the $\theta_w$ gradient has also been identified in the observational study by Neiman et al. (1993)—see their Fig. 13. The present multiple structure may be related to the secondary cold fronts observed by Kreitzberg (1968) and to the secondary surges of low-$\theta_w$ air behind the system observed by Hobbs et al. (1975).
Figure 4. Surface and cloud analysis for 18 UTC 27 April. See text for details.

Figure 5. Visualization of the broad three-dimensional pattern of airflow in and near the cloud head at 18 UTC 27 April, obtained from relative-flow analyses in three isentropic surfaces. Thick solid and thick open streamlines show saturated flows in the 12 °C and 8 °C \( \theta_e \)-surfaces, respectively. Thin solid streamlines show the dry flow in the 25 °C \( \theta \)-surface. Dashed and dotted lines show heights of these three surfaces at 1 km intervals. Stippling represents saturated flows and dashed shading denotes relative humidity less than 30% within the corresponding isentropic surfaces. The area of this figure corresponds exactly to that of Fig. 4. Other details in text.
Figure 6. Distribution of $\theta_e$ at 950 mb for 18 UTC 27 April, derived from a subjective analysis of the drop-windsondes. Isopleths are labelled with large numbers (°C). The small numbers show the $\theta_e$-values for individual soundings displaced according to the system velocity. The tip of the warm sector/low centre was determined by a ship observation. The analysis is reliable in detail only close to the plotted data-points; fine structure seen in the satellite imagery suggests there may have been a somewhat more fragmented $\theta_e$ pattern in some other places.

Figure 7. Meteosat visible image for 1730 UTC 27 April, showing the well-defined clearance of cloud immediately behind CFI (thin black strip orientated SW–NE and passing through 15°W 45°N). The inner frame shows the area covered by Figs. 3, 4, 5, 6, 8 and 9.
In the satellite data, cloud lines associated with the cold fronts can be seen in both the visible (Fig. 7) and infra-red (Fig. 8) images. CF1 is identified particularly clearly in Fig. 7 by the sharply-defined dark strip associated with a clearance of cloud immediately behind the surface front. Figure 8 shows that the band of cloud immediately ahead of CF1 consisted of low clouds with tops ranging from +5 °C (at the blue to green transition) to −5 °C (at the medium-green to light-green transition). For a $\theta_w$ of 12 °C this corresponds to tops between 1.4 and 3.1 km. This cloud band was observed from the C-130 aircraft as a rather smooth-topped line of stratocumulus (rope cloud). Slightly higher (3 km) cloud tops within this line can be seen (light green in Fig. 8) to have originated close to the surface position of CF1 and then to have extended forwards at a small angle (about 20°) to the front; these features were parallel to the relative airflow at their tops. (See $\theta = 25 °C$ flow overtaking the surface front in Fig. 5.) Cold front CF1 extended to the position of the low centre and here the cloud tops rose abruptly to a maximum of 6.5 km (red in Fig. 8); this region of mid-level convection is discussed in detail in section 4(c).

The second cold front, CF2, was rather less distinct in the imagery than CF1. At visible wavelengths it is seen as a ragged narrow band of cloud. Along part of its length there is another thin line of cloud just ahead of it. To the north-east, these thin lines of cloud merge into the tip of the cloud head, part of which is itself considered to be due to moist air flowing slantwise above the low-$\theta_w$ air associated with CF2. The irregular nature of the cloud band along CF2 compared with the smooth texture of that ahead of CF1 is an indication of the more intermittent nature of the convection, due perhaps to the dryness of the air overrunning the warm sea. The convection here was strongly suppressed by the descending dry air and the associated cloud tops were no higher than 1.5 km (dark-green band along CF2 in Fig. 8). Behind CF2, the linear nature of the cold-frontal cloud gave way to irregular open-cellular convection typical of subsided arctic airflows over a relatively warm sea. The sea surface temperatures in the overall frontal region were between 11 and 13 °C. Although this led to the potentially unstable boundary layer behind CF1, in the warm sector ahead of CF1 there was a near-neutrally-stable boundary layer where dry- and wet-bulb temperatures, at 12 to 13 °C, were about equal to the sea surface temperature. As discussed in section 4(b), the near-neutral stability here is an indication of the effectiveness of the convection in creating a rather well-mixed boundary layer.

(d) Visualization of the flows associated with CF1, CF2 and the cloud head

The flow of cold air associated with CF1 and CF2 was interrupted by the interleaving of the flow of moister air circulating rearwards in the south-western tip of the cloud head. Before examining the two cold-frontal flows we shall first consider this cloud-head flow, as represented in Fig. 9(a). To establish a visual link between this figure and other parts of Fig. 9 (and Fig. 8) we have outlined the region of the cloud head where the moist layer (RH > 90%), derived from the dropsondes, extended above 1.5 km. Figure 8 shows that both this contour and the 3 km contour are consistent with the satellite cloud-top pattern (except to the south-west of Brittany where the cloud-head moisture boundary was obscured by much higher cloud associated with the polar-front cloud band).

We now consider the cloud-head flow in detail. On the right side of Fig. 9(a) the dot-dashed contour shows the location of a shallow layer of rather dry air (RH < 90%) whose base was at 0.5 km. This corresponded to the air (with $\theta = 13 °C$) within the CCB ahead of the warm front that subsequently entered the cloud head. The streamlines in Fig. 9(a) show this flow rising towards the north-west, with its RH increasing to 100%. The subsequent saturated flow is plotted within the $\theta_w = 7 °C$ surface which shows the CCB flow eventually emerging at the tip of the cloud head at between 1 and 2 km. This is the flow that was interleaved between the cold flows approaching cold fronts CF1 and CF2.
Figure 8. Meteosat infra-red false-colour image for 18 UTC 27 April, with surface frontal positions taken from Fig. 6. Cloud-top temperatures (°C) are given by the colour code along the bottom. Also shown are isopleths of the depth of the moist (90% relative humidity) layer deduced from the dropwindsondes.

The airflows that penetrated to low levels just behind CF1 and CF2 are represented in Figs. 9(b) and (c) by the relative flows in the \( \theta = 20 \, ^\circ\text{C} \) and \( 10 \, ^\circ\text{C} \) surfaces respectively. The \( \theta = 20 \, ^\circ\text{C} \) flow in Fig. 9(b) can be seen descending steeply from the north-north-west, the eastern part of it overrunning the tip of the cloud head. By the time this flow reached the position of CF1 it was at 1.5 km, tending to overrun CF1 and turning towards the north-east. Mixing processes are believed to have brought some of this air down to the sea surface and, indeed, the dropwindsondes showed that in the diagonally hatched area in Fig. 9(b) the RH = 90% contour dropped as low as 0.5 km, indicating strong downward penetration of the cold dry air along this part of CF1. In Fig. 9(c) the \( \theta = 10 \, ^\circ\text{C} \) flow behind CF2 is near the boundary layer and so the analysis can be relied upon in only a semi-quantitative sense; nevertheless it is clear that the \( \theta = 10 \, ^\circ\text{C} \) air was descending and may have reached the surface just behind CF2. The eastern part of the flow can be seen undercutting the tip of the cloud head.

4. EVOLUTION OF THE COLD FRONTS

(a) The dry intrusions associated with CF1 and CF2

The satellite water-vapour (WV) imagery, indicating the pattern of water vapour in the middle and upper troposphere, showed two closely-spaced dark zones due to two dry intrusions which appeared to be associated with the two surface cold fronts. Figure 10 shows a 3-hourly sequence of the WV imagery, with the estimated positions of surface cold fronts CF1 and CF2 superimposed. The first dry-intrusion straddles CF1, consistent with the overrunning shown in Fig. 9(b). The second dry-intrusion lies immediately behind CF2. The dark zone associated with the latter dry-intrusion was growing in extent, presumably
Figure 9. System-relative flows at 18 UTC 27 April, derived from dropwindsondes. The flows represent: (a) moist air in the $\theta_v = 7 \, ^\circ C$ surface within the cloud head, (b) dry air in the $\theta = 20 \, ^\circ C$ surface behind and slightly overrunning CF1 and (c) dry air in the $\theta = 10 \, ^\circ C$ surface behind CF2. Part of the flow in (a) intrudes between the flows in (b) and (c). Dashed lines show the heights of these surfaces. Dotted lines show relative humidity. Also shown are the position of the cyclone centre (L) and the extent of moist air (relative humidity $> 90\%$) above 1.5 km associated with the cloud head (solid contour with stippling). The 0.5 km base of a shallow layer of dry air beneath the warm-frontal zone is denoted by the dot-dashed isopleth in (a). Another region, along CF1, where dry air penetrated downwards as far as 0.5 km is shown in (b) by hatching.
in association with descent over a deep layer. However, the dark zone associated with the first dry-intrusion diminished in extent, becoming quite narrow by the end of the period. This is thought to have been partly because of stretching deformation and partly because it was within the region where the dry-intrusion air was turning towards the north-east and ascending above a moist zone (see also Fig. 5). Such a behaviour is consistent with the evolution of CF1 as diagnosed below.

(b) Transition of CF1 from a surface cold front to an upper cold front

Figure 11 shows part of the surface analysis and associated cloud-head for two occasions 9 hours apart. The first analysis corresponds to the dropsonde analysis discussed in section 3. The second analysis, based mainly on reports over France and the United Kingdom, shows an interesting change in structure: whereas CF1 and CF2 were both diagnosed as surface fronts at 18 UTC, CF1 had become purely an upper front by the later time. No corresponding wind change or temperature change occurred at the surface with the passage of CF1 over stations such as Jersey. Indeed at the time of the passage of CF1 over Jersey,
and stations in Brittany, the surface temperature and dew point were still increasing and continued to increase slightly for some time. Meanwhile CF2, which previously had been more than 100 km behind the warmest surface air, had come right up to the edge of the warm sector to become the main surface cold front.

Evidently, the two cold fronts were travelling eastwards faster than the rest of the system, where we define the rest of the system as being the combination of the cloud head, cyclone centre and the belt of low-level 'warm sector' air with highest $\theta_w$. The motion of the cold fronts was related to the higher velocity of the air in the overrunning dry-intrusions (Fig. 9(b)). The warm sector was situated within a region of ascent ahead of the trough and, as the dry intrusions approached, the formerly descending dry air ascended and overran the high-$\theta_w$ air in the warm sector. Thus by 03 UTC 28 April the region between CF1 and CF2 was characterized by a shallow moist zone (SMZ) overlain by dry air as in the split-front model of Browning and Monk (1982). Infra-red satellite imagery showed that the cloud tops were very low (between 1 and 3 km) in the SMZ. At this stage the reader will recognize similarities to the simple conceptual model in Fig. 1, where the leading edge of the overrunning dry-intrusion is depicted as an upper cold front (open cold-front symbols).
This is the feature described as an upper cold front by Browning and Monk (1982), and is probably related to the features described as a prefrontal surge by Kreitzberg (1968), a cold front aloft by Hobbs et al. (1990) and an upper-level humidity front by Mass and Schultz (1993).

The fact that CF1 lost its identity close to the surface as it migrated through the system is intriguing. The loss of identity at the surface occurred more or less as it entered the pre-trough region of large-scale ascent. The track of the cyclone system was such that air arriving in the boundary layer behind it was potentially unstable. As a fresh pulse of cold air corresponding to CF1 approached, there would have been an increase in convective mixing. But, because of the large-scale subsidence, the mixing was only partial and pockets of low-\(\theta_w\) air could be detected penetrating to the surface just behind the front. However, as soon as the pulse of colder air encountered a region where the large-scale ascent was sufficient to release most of the potential instability, the mixing of the boundary layer proceeded more effectively. This led to an SMZ in which the air was closer to being neutrally stable: it tended to be fairly close to saturation and \(\theta_w\) was nearly constant with height at a value close to that of the temperature of the sea surface. It is possible that this same mixing was responsible for eroding away the lower part of CF1 at this later stage. Thus, although CF1 was detectable as a surface feature at one stage and later only as an upper-level feature, this was not due to cold surface-based air riding over the warm air but, rather, to low-\(\theta_w\) air from aloft no longer penetrating relatively unmixed to the surface.

(c) Middle-level convection along the upper cold front

Figure 12 shows hourly plots of CF1 and the clouds along it as it evolved from being a surface front to a purely upper front. Most of the cloud tops were below 3 km (warmer than \(-5^\circ\)C) but near the cyclone centre, as time progressed, they rose above 6.5 km (colder than \(-30^\circ\)C). The higher convective cells formed in a short line, each cell developing at the southern end and dissipating after a few hours at the northern end. A radiosonde was released at 23 UTC from Brest at the leading edge of the SMZ close to the position of the upper cold front, CF1. It shows significant middle-tropospheric potential (and actual) instability (Fig. 13), consistent with the occurrence of the cold-topped convective clouds that were observed along CF1 a few tens of kilometres to the north of Brest (Fig. 12). The cold-topped convective clouds occurred only within 100 km of the surface cyclone; presumably the lifting required to trigger the mid-level instability did not extend beyond this region.

In order to keep Fig. 12 simple, we have not shown the high cloud tops that developed before 16 UTC and were located at the northern end of CF1 for a while after 16 UTC. They behaved just like the other convective cells described above, but further insight into their origin is provided by the dropsondes that were released between about 1430 and 1530 UTC during Run 3 (see Fig. 3). This run was almost parallel to CF1 and a little ahead of it. Its position with respect to the near-surface fronts is shown in Fig. 6 by the line of numbers extending from the underlined 8 ahead of the warm front to the underlined 12 within the warm sector just ahead of CF1. A vertical cross-section of \(\theta_w\) along this line is shown in Fig. 14. The most obvious feature of Fig. 14 is the inclined warm-frontal zone reaching the surface halfway along the section. However, Fig. 14 also shows a feature that is related to CF1 and the associated mid-level convection: this is the layer of low-\(\theta_w\) air between 2 and 3 km (labelled CC in Fig. 14) intruding above the high-\(\theta_w\) warm-sector air. This intrusion of low-\(\theta_w\) air corresponds to the dry air in the 25 \(^\circ\)C \(\theta\)-surface that was overrunning the surface position of CF1 at the time depicted in Fig. 5. It produced a shallow layer of potential instability which was realized in the region of ascent close to the cyclone centre where this air began to overrun the warm-frontal zone. The resulting mid-level convection
Figure 12. Evolution of convective clouds associated with CF1 that developed after 16 UTC 27 April. The successive hourly positions of cloud tops have been derived from Meteosat infra-red imagery. Stippling and hatching denotes tops colder than −5 °C and −30 °C respectively. Short arrows show tracks of individual cloud-elements. Hourly positions of CF1 are also indicated. The track of the surface cyclone was between XX. B denotes the location of the Brest ascent plotted in Fig. 13.

Figure 13. Tephigram showing temperature and dewpoint sounding from Brest at 23 UTC 27 April. Also shown are winds (in knots). See Fig. 12 for location.
Figure 14. Cross-section along Run 3 within the warm sector just ahead of CF1 (see location in Figs. 3 and 6), showing pattern of $\theta_v$ isopleths at intervals of 0.5 degC derived objectively from the dropwindsondes. Locations of the 10 sondes are shown at the top of the diagram. The bold solid line demarcates the region of strong convergence ($> 4 \times 10^{-5} \text{s}^{-1}$) associated with the warm-frontal zone. The dashed line labelled CC shows the intrusion of relatively dry, low-$\theta_v$, air overrunning the moist warm-sector air ahead of CF1. The schematic cloud-turret shows the location of middle-level convection whose associated cloud tops and precipitation were observed by satellite and aircraft respectively.

is depicted by the schematic cloud-turret in Fig. 14, which extended to the aircraft’s flight level and was evident on the aircraft’s radar display as an area of upper-level precipitation 30 km wide.

Figure 15 shows the weather-radar rainfall-pattern superimposed on the corresponding Meteosat infra-red image at a late stage in the evolution of CF1. The convective clouds associated with the upper cold front, CF1, gave rise to the moderate rain showers (dark blue) seen extending from Cherbourg to central southern England. The rain showers were weakening at this time as CF1 travelled eastwards away from the cyclone centre. This is consistent with the lowering (warming) of tops of the deepest convection from 22 UTC 27 April onwards as shown in Fig. 12.

5. Vertical structure of the dry intrusions penetrating into the low troposphere near CF1 and CP2

We shall now focus on detailed analyses of dropsondings during Run 5 which provided a cross-section roughly north–south through the south-western tip of the cloud head (see location of Run 5 in Fig. 3). There were two levels of complexity in the observed
structure: the primary double structure corresponding to the two dry intrusions and associated cold fronts CF1 and CF2 (described in section 5(a) below), and some coherent structures occurring on an even finer scale (described in section 5(b)).

(a) The primary features of the vertical structure associated with CF1 and CF2

The two dry intrusions and the two associated cold fronts show up well in the fields of relative humidity and $\theta_w$ reproduced in Figs. 16 and 17 respectively. Although there is much fine structure (discussed later) this does not obscure the two main thrusts of dry low-$\theta_w$ air that are situated on different sides of the tip of the cloud head (see large arrows labelled 1 and 2 in Fig. 17). The features of the $\theta_w$ pattern that relate most clearly to the double frontal structure are:

- **Dry Intrusion 1.** This corresponds to the overrunning flow of air above the cloud head, with a region of $\theta_w$ between 9 and 11 °C descending towards the surface in advance of the cloud head but behind the cloud (C1) that was associated with CF1. This flow was subsaturated and corresponds roughly to the air with $\theta = 20$ °C as analysed in Fig. 9(b).
- **Dry Intrusion 2.** This corresponds to the flow of air with low $\theta_w$, between 4 and 6 °C, extending to the surface behind CF2 which was itself marked by a locally sharp $\theta_w$-
Figure 16. Cross-section along Run 5 (see location in Fig. 3) showing isopleths of relative humidity derived subjectively from dropwindsondes, superimposed upon the outline of cloud tops derived from infra-red imagery. Locations of the 17 sondes are shown at the top of the diagram.

gradient. This airflow was initially subsaturated and corresponds roughly to the flow of air with $\theta = 10$ °C that was analysed in Fig. 9(c). (The presence of virtually saturated air all the way to the surface beneath the cloud head in Fig. 16, especially near 200 km, may be an artefact of the humidity sensor; it seems more likely that subsaturated dry-intrusion air, with correspondingly lower $\theta_w$ values, will have undercut the cloud head all the way to the surface position of CF2.)

- **Cloud Head.** Air with $\theta_w$ between 6 and 9 °C was associated with the cloud head itself. This corresponds to the flow of air that was analysed in Fig. 9(a) and Fig. 5. Much of the cloud material within the part of the cloud head observed in Run 5 was formed by ascent that occurred in the main part of the cloud head to the north-east of Run 5. As discussed in section 3(b), it is not clear exactly what the vertical air motion was in the top of the cloud head. Probably the best way of thinking of it is that this was a cloudy region of near-zero vertical velocity slicing into the region of otherwise descending air associated with the dry intrusions.

Figure 18 shows that the rear edge of Dry Intrusion 1, where dry low-$\theta_w$ air was plunging down just ahead of the cloud head, was characterized by an inclined layer of strong vertical shear in the component of the wind parallel to the cold front. The simple geometry of the layer implies that this component of the shear also corresponded to a horizontal shear in a cyclonic sense. Below 3 km the shear was split into several thinner
Figure 17. As Fig. 16 but showing isopleths of $\theta_v$ instead of relative humidity. The arrows labelled 1 and 2 draw attention to the two dry intrusions, with low $\theta_v$, associated with Cold Fronts 1 and 2 (the arrows are schematic, the extent of the dry intrusions being much larger than them).

layers but this is fine structure which we shall refrain from discussing until the next subsection. All the regions of high shear tended to be associated with high static stability but the primary shear-layer at the top of the cloud head was characterized by an especially well-defined strip of high static stability between the two positions marked X in Fig. 18. The combination of strong cyclonic shear and high static stability suggests that the layer may have been a region of high potential vorticity (PV). A parallel argument suggests that the neighbouring layer may have been characterized by fairly low PV.

Figure 19 shows a cross-section of PV derived objectively with the assumption of two dimensional (2-D) symmetry along the direction of the cold-frontal zone. This analysis suffers from two deficiencies. Firstly, the 2-D assumption is not very accurate so close to the cyclone centre (particularly since Run 5 was not perpendicular to the frontal zone) and so it is not possible to estimate PV reliably without simultaneous measurements in the perpendicular plane. Secondly, the fine-scale nature of the structure makes it difficult to determine differentiated quantities; the transformed geostrophic-coordinate analysis of Desroziers and Lafore (1993) might have improved the analysis, although still only within a 2-D assumption. Notwithstanding these limitations, there is a strong suggestion of high PV within the primary shear-layer at the top of the cloud head due to descent of dry-intrusion air from the upper troposphere (if not from the stratosphere itself). Where this air encountered the boundary layer it probably forced vertical circulations, which may
account for the upward extension of moist air into the cloud head and the other fine-scale structure described next.

(b) Additional fine-scale structure associated with the dry intrusions

The subjectively analysed fields in Figs. 16, 17 and 18 show a considerable amount of fine-scale structure. (Note that subjective analyses have been employed because objective analyses were less successful in resolving the fine structure; confidence in the subjective analyses is high, given both the level of agreement between near-neighbour soundings and the close correspondence between patterns derived from independent measurements.) The fine-scale sub-structure is seen as multiple, inclined, thin fingers. Corresponding fingers are evident in another vertical section farther west (Run 6, not shown) and so the fingers were evidently sections through inclined laminae. These laminae show up especially well in the pattern of RH depicted in Fig. 16 (see the multiple dashed lines). A typical vertical spacing between the axes of dry and moist laminae within a given dry-intrusion is 0.5 km. The component of their slopes within the observed section was about 1 in 100. Allowing for the fact that the line of soundings cut obliquely across the cold frontal zone, the true slope of the laminae was about 1 in 60 and their horizontal spacing 60 km.

One particularly well-defined dry lamina, associated with the primary shear-layer just above the cloud head, extended for over 200 km in the horizontal and 2 km in the vertical.
It was resolved in four consecutive dropsonde-profiles, two of which are reproduced in Fig. 20. The overlying moist lamina in Fig. 20 was just as extensive as its dry counterpart and although its maximum RH often exceeded 80% it did not (quite) reach saturation above 1 km. This is consistent with the infra-red imagery which showed that, apart from a narrow line of cloud below 1 km, this moist lamina was situated in a cloud-free region between the cloud band associated with CF1 (labelled C1 in Fig. 16) and the main cloud-head. Judging from the consistency of the overall fit between the two sets of data, we believe that the registration between the dropsondes and the imagery was good to within 10 km.

Figure 18 shows that the primary shear-layer, which extended from above 5 km, downwards ahead of the cloud head, split into several thinner laminae below 3 km. The three best-defined laminae in Fig. 18 are seen to correspond closely to the independently-measured RH laminae in Fig. 16. The dashed curve in Fig. 18 shows that these laminae lay within a region where the component of the wind resolved along the direction of travel of the large-scale system was close to the system velocity. We discuss the possible significance of this observation later in this section.

Mesoscale banded structure is common in frontal systems and is a topic of much theoretical interest. Recent research has tended to favour symmetric instability (SI) or conditional symmetrical instability (CSI) (Bennetts and Hoskins 1979) as the cause of such
Figure 20. Tephigrams and winds (knots) for two adjacent dropwindsonde profiles along Run 5, showing the moist and dry laminae just above the moist air associated with the south-western tip of the cloud head. Their locations are shown in Fig. 3: (a) and (b) reached mid-flight at 1700 and 1706 UTC respectively.

structure, although few highly-resolved observations are available to provide definitive confirmation (Thorpe and Clough 1991). According to that view, the main discriminant is the occurrence of negative PV, or negative equivalent PV (hereafter referred to as EPV, as defined by Thorpe and Clough) in saturated regions. Unfortunately, the pattern of EPV (in Fig. 21) is even less well-defined than that of PV (in Fig. 19) and the caveats in the above discussion of PV apply even more strongly for EPV. Nevertheless, there are, perhaps, some features in Fig. 21 that may be significant: one is the tendency for bandedness to occur
across the frontal zone, i.e. sloping upwards to the right in Fig. 21, as observed previously by Thorpe and Clough (1991) in the FRONTS 87 experiment. This behaviour is unlike that found in idealized-model simulations of CSI (e.g. Persson and Warner 1993), although these were necessarily carried out under a simple 2-D assumption. It is also evident that most of the regions of strongly negative EPV in Fig. 21 are in the dry air and are unlikely to have been realised in the prevailing conditions of descent. There are small regions of negative EPV in parts of the cloudy regions but there is no strong indication that CSI (or SI) is a primary driving force for the mesoscale circulations in this case or that it alone can provide an adequate explanation of the observed multiple laminae.

The evidence is unfortunately limited, but the hypothesis of descent of pre-existing PV anomalies may yet offer a plausible explanation for the occurrence of the multiple laminae. We are accustomed to thinking of dry intrusions bringing down high-PV air from tropopause level (e.g. Browning and Reynolds 1994) and, indeed, we believe this happened within the primary-shear layer at the top of the cloud head discussed in section 5(a). However, this is only one side of the coin: the upper-tropospheric part of a jet stream often possesses low, and sometimes even negative, PV, generated evidently by previous latent-heat-release upstream (Thorpe and Clough 1991), and this too will have been brought down. According to semi-geostrophic theory, the response of the vertical velocity to geostrophic forcing is given by an appropriate omega equation, in which the role of PV resembles that of static stability in quasi-geostrophic theory (Hoskins and Draghici 1977). Thus it should be expected dynamically that large-scale forcing intense enough to cause significant descent
of stratospheric air with high PV will cause even greater descent of adjacent low-PV air. Figure 19 shows that there were patches of negative PV at 4 km, with an inclined lamina of relatively low PV (PV < 0.5 PVU)* extending downwards just above the line XX. A value of just under 0.5 PVU is not particularly low by tropospheric standards; however, because this lamina was a shallow feature sandwiched between laminae with high PV, its true PV-value may not have been resolved even with the high density of soundings available in this study.

The near-critical location of the multiple laminae with respect to the frontal wave, demonstrated above, suggests that critical-layer behaviour may have been playing a role in determining the mesoscale structure embedded within the system. Critical-layer effects occur in parcels of air that move at the same velocity as a wave in which a forcing process operates; such parcels are subjected to protracted forcing and will thus show the results of forcing more intensely than those at other locations in the fluid. We are not aware of any observational studies highlighting such processes, but their importance has been noted by Hoskins and West (1979) and Davies et al. (1991) in the context of theoretical studies of frontogenesis. These authors considered the formation of fronts in dry 3-D semi-geostrophic models of baroclinic wave development. Hoskins and West (1979) demonstrated that the greatest frontogenesis occurred where parcel trajectories stayed longer within the region of intense frontogenesis, while Davies et al. (1991) showed that frontogenesis in different basic-state flows was most pronounced in regions of the fluid moving close to the phase velocity of the unstable baroclinic mode. The forcing process in the models was purely-ageostrophic motion and this appears likely to have been a major factor in the case considered here. However, the proximity to cloud in the present case suggests that latent-heat release may have provided an important amplifying effect. In general, forcing may be due to diabatic processes if these have a fixed or slowly-varying phase relative to the wave (Lin and Chun 1991). They may thus be expected to 'etch' out a pattern with scale and intensity determined not only by the extent and strength of forcing but also by the duration implied by the system-relative velocity. Since the speed of a wave changes as it develops, the critical region will differ with time, probably tending to descend as a wave decelerates during growth.

In summary, low EPV is thought to have been important for the development of the fine structure in this case, but the viewpoint adopted here is different from the conventional view that such structure is determined mainly by local instabilities such as CSI. Our suggestion is, firstly, that the externally imposed baroclinic-wave phase-speed is a primary influence and, secondly, that the importance of local stability may arise in determining the amplitude of response to an imposed forcing (such as divergence of the Q vector) rather than by implying a spontaneously unstable growth. We believe that it is significant that much of the fine-scale structure in the present study occurred at heights between 1 and 3 km, close to the edge of the moist boundary layer. It is therefore possible that latent heating had some influence on the fine-scale structure, though this influence may have been exercised primarily by determining the local stability, thereby increasing the amplitude of the secondary response to the forcing perturbation, as suggested by Thorpe and Emanuel (1985).

6. Conclusions

We have presented a case study of a moderately rapidly developing frontal cyclone which displayed a characteristic structure comprising cloud head and dry intrusion—in

* 1 unit of potential vorticity (PVU) is $10^{-6} m^2 s^{-1} K kg^{-1}$
this case two dry intrusions in close succession. Most frontal cyclones contain at least one dry intrusion, which is the specific name given to the region of dry, low-$\theta_w$, air that enters the cyclone system from the rear following a period of descent from near a tropopause fold. In this study we have used an unusually dense array of soundings to focus on the structure of the dry intrusions and on the way in which they interacted with the moist air beneath them in the vicinity of the cloud head.

The double structure of the dry intrusion was manifested aloft as two distinct dry slots (dark zones) in the satellite water-vapour imagery and, at the surface, as two cold fronts 100 to 200 km apart. The dry intrusions and associated cold fronts travelled faster than the cyclone system. At one stage both cold fronts were surface features. Over the period of the study, the leading cold front advanced ahead of the cyclone centre; however, the associated low-$\theta_w$ air did not displace the warm-sector air and lead to a process of occlusion as in the classical Norwegian model (Bjerknes and Solberg 1922). Instead, the lower portion appeared to be eroded by boundary-layer mixing and the upper portion became decoupled from the surface, riding over the shallow moist zone of high-$\theta_w$ air in the warm sector which was feeding into the cloud head to the north of the cyclone centre. The first cold front thus became an upper front (strictly an upper-level cold $\theta_w$-front), and the second cold front then took its place as the new surface boundary of the warm sector. This is evidently one of the ways in which the split-front structure in the model of Browning and Monk (1982) can develop. Presumably a similar process could occur even in the absence of a double dry-intrusion, in association with different parts of a single dry-intrusion, i.e. when just the leading edge overruns the warm sector.

The surface cold fronts in the present study were characterized by thin lines of shallow stratocumulus that extended southwestwards from the cyclone centre. They were kata-fronts (Bergeron 1937) and over much of their length they produced hardly any precipitation. When the first cold front rode over the shallow moist zone (warm sector) and developed into a purely upper front, a line of mid-level convective shower cells formed along it. The showers were confined to a short line within 100 km of the cyclone centre where there was sufficient lifting above the warm front to realize the potential instability.

The main precipitation region of the cyclone was associated with the cloud head located mainly to the north-east of the cold fronts. The cloud head was predominantly due to ascent of cold-conveyor-belt air, though there is evidence, not shown here, that ascent of air from the base of the warm conveyor-belt (W2) also contributed to the cloud head as in the model of Browning and Roberts (1994). The stratiform cloud deck that composed the present cloud head reached above 6 km over a large area but the main focus of this study was on the region near the south-western tip of the cloud head where the cloud top was below 3 km. Here, cloud, which had formed during earlier ascent within the main part of the cloud head, was travelling south-westwards relative to the system and becoming interleaved between the two dry intrusions.

The region where this interleaving occurred, south-west of the cyclone centre, was characterized by considerable fine structure. Inclined laminae characterized by strong shear and cyclonic vorticity, associated with the dry intrusions, split into multiple inclined laminae near the top of the boundary layer at heights of between 1 and 3 km. Our detailed observations provide a glimpse of the ordered and highly laminar interactions that occurred where descending air within dry intrusions encountered the moist boundary layer. The multiple layers were revealed not only in the kinematic fields but also in the independently-measured fields of RH and $\theta_w$. This structure was caused by the distortion of the boundary-layer top by localized shears and associated differential displacements in the vertical. Individual moist and dry laminae were inclined at an angle of 1 in 60, with a vertical spacing between axes of adjacent moist and dry laminae of about 0.5 km.
The moist laminae corresponded to air from within, or near, the boundary layer, and the dry laminae corresponded to the dry-intrusion air penetrating slantwise into the boundary layer. It was pointed out that this structure occurs in air travelling close to the speed of the baroclinic wave, and we have thus suggested that this mesoscale structure may result from a near-critical-level response to forcing rather than a spontaneous mesoscale instability such as CSI. However, adequate discrimination between such mechanisms would require 3-D observations of high resolution to define the mesoscale PV and EPV fields. Even the present high-quality data-set does not meet the stringent requirements of near simultaneity for such an analysis.

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REFERENCES

Hobbs, P. V., Locatelli, J. D. and Martin, J. E. 1990 Cold fronts aloft and the forecasting of precipitation and severe weather east of the Rocky Mountains. Weather Forecasting, 5, 613–626
Hoskins, B. J. and Draghici, I. 1977 The forcing of vertical motion according to the semi-geostrophic equations and in an isentropic co-ordinate model. J. Atmos. Sci., 34, 1859–1867


