Mesoscale dynamics of cold fronts: Structures described by dropsondings in FRONTS 87

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SUMMARY

The FRONTS 87 project was a European experiment to make mesoscale observations to deduce the dynamics of active cold fronts approaching north-west Europe, with a minimum of orographic influences. In this paper data from dropsondings made by the Meteorological Office's C-130 research aircraft are analysed. The aircraft flew a pattern with 4 or 5 runs oriented approximately at right angles to the front, but with each run displaced in the along-front direction by about 100 km. The dropsondes gave soundings to a height of approximately 7 km with a cross-frontal spacing of 20 km at best and 60 km on average. Between 30 and 50 soundings were made in each event, over an area of 500 x 500 km². These data are unique in their mesoscale resolution over a synoptic-scale region; this was made possible by the ability to track up to 5 sondes in the air at any time.

Several new aspects of cold-frontal dynamics have been identified from these data; the approach here is to treat each run as an instantaneous cross-frontal section. The discussion is centred on the hypotheses postulated before the experiment; these were intended to allow verification or refutation of aspects of models of frontal dynamics such as those embodied in semi-geostrophic theory. A minimal and purely objective analysis has been performed on the data presented here, but the high resolution of the dropsonde observations permits evaluation of differentiated quantities such as potential vorticity (PV), with some confidence. Issues raised here include the degree of thermal-wind balance, the structure of conserved variables such as equivalent potential temperature and absolute momentum, the cross-frontal circulation and the role of diabatic processes such as the evaporation of snow, and the potential vorticity structure on the mesoscale and its interpretation. The intention is to present an integrated view of the dynamical structure of fronts in the light of theoretical rather than conceptual or airflow models.

1. INTRODUCTION

The Mesoscale Frontal Dynamics Project (MFDP) is a long-term programme of research into the dynamics of active fronts, with participation from groups in the United Kingdom and France. The project originated from the need to bring together three distinct areas of frontal research: conceptual models, theoretical ideas, and mesoscale observations. Observational technologies now available should allow the conceptual and theoretical models to be evaluated critically. The first field-observation project within the MFDP combining the expertise of these groups was held from October 1987 to January 1988 and is called FRONTS 87. The individual groups had conducted separate field programmes previously such as that described by Testud et al. (1980), FRONTS 84 (Lemaitre and Scialom 1990), Project Scillonia (Roach and Hardman 1975), and that described by Bennetts and Ryder (1985). A series of internal project documents have described the aims of the project (Browning and Testud 1986), the theoretical aspects (Thorpe et al. 1987), the field observation details (Lunnon and Lemaitre 1987), and preliminary results (Machin 1988; Shutts 1988). In addition Browning et al. (1986) and Clough and Testud (1988) provide an overview of aspects of the project. Here we will survey the scope of FRONTS 87 with particular reference to the range of observations made and the theoretical hypotheses to be tested with these data; further details can be found within the previously listed references.

To set the scene we reiterate the specific scientific objectives of FRONTS 87 as described by Browning and Testud (1986).

(1) To obtain an improved dynamical understanding of synoptic, mesoscale and smaller scale interactions within systems containing cold, especially active cold, fronts.
(2) To acquire mesoscale data-sets and use them for the further development of numerical models and the parametrizations in them.

(3) To describe the structure and evolution of mesoscale features in such situations, within the full synoptic context, and to derive conceptual models of value in very-short-range forecasting (0–12 h).

A focal point of the project is to attempt to describe and understand interactions between the various scales: to this end Browning and Testud (1986) define the following scales. The synoptic scale has motion governed by quasi-geostrophic dynamics on the horizontal scale related to the Rossby radius of deformation for depth scale \( H \), \( L = NH/f \) or about 1000 km where \( N \) is the Brunt-Väisälä frequency and \( f \) is the Coriolis parameter. Observations at a spacing of about 300 km are required to define such structures. The mesoscale is, it is assumed, the scale of observable distinct structures at fronts such as frontal rainbands and sloping ascent zones. This scale can be defined as that on which ageostrophic advection is important so that \( L = u/f \) or about 100 km, where \( u \) is an appropriate horizontal wind speed. To define these systems adequately observations are needed at spacings of 20–30 km. Finally the small scale of cloud processes and gravity waves will have \( L = H = u/N \) or 1 km. Hydrostatic balance no longer applies and observations are needed with spacings of a few hundred metres to define processes on this scale.

A wide range of observing technology was used in FRONTS 87, including aircraft (observations \textit{in situ} and by dropsonde), radar (including dual-Doppler), and routine radiosondes. Each of these systems was used to look at one or more of the three scales described above. In this paper we concentrate only on the aircraft dropsonde data as these were specifically designed to define the mesoscale structure over a region which was large enough to capture the essence of the synoptic-scale forcing. The Meteorological Office’s C-130 aircraft produced dropsonde profiles of temperature, humidity, and winds below about 8 km with horizontal spacing down to about 20 km. Lunnon and Lemaitre (1987) describe the optimal flight plan for the deployment of the dropsondes: it corresponds, in a coordinate system moving with the front, to a pattern of five runs in the across-front direction mapping out a hexagonal region of dimensions about 500 \times 500 km\(^2\). The aim was to provide a relatively uniform coverage over the region but with increased resolution in the across-front direction by using closely spaced soundings in the central run. In later flights only four runs were used, these being extended the better to observe the upper front. The flight plan reveals an important purpose of the observations, namely an attempt to estimate the along-front variations. Each of the cross-front runs took 1–2 hours to fly thus producing a dropsonde launch interval, for a 20 km resolution, of about 3 min (the fall period of sondes was typically 12 min). The total time between the first and last sonde in the pattern was about 5 hours. The mixture of spatial and temporal dimensions in these observations is a serious issue but is, of course, typical of aircraft-derived data. The inherent supposition is that at least for each of the 1-hour runs the front remains steady; intercomparison between the runs supposes, rather less justifiably, a period over which there is steadiness of the front. However it should be emphasized that these data represent unique observations of fronts at high spatial and temporal resolution over a large-scale domain. Other published frontal cross-sections involve either synoptic data or long time-series at individual stations. These problems of data interpretation are inherent in most mesoscale observational strategies currently employed.

It is the aim in this paper to present mesoscale structures common to most if not all of the cases observed. It is not our intention to provide a comprehensive picture of any
given front, nor do we wish to use information that is extraneous to the data. In contrast, in subsequent papers we will present objectively analysed fields in three dimensions derived from dropsonde and other data. Also no reference will be made here to weather-prediction-model forecasts; although again these will be the subject of future publications. The unique high spatial resolution of the dropsonde data is being exploited here to examine, without undue prejudice, whether mesoscale structures defined in the theoretical models, or ones not so defined, are evident in the 'raw' data. The philosophy being employed is that each front or even each aircraft run has, by itself, less significance than structures defined by several runs, be they from a given front or from other fronts. Of course some particular runs and fronts are more useful for demonstrating some points than others.

2. THEORETICAL HYPOTHESES OF FRONTS 87

Thorpe et al. (1987) have reviewed current theories of frontal dynamics and formulated key scientific hypotheses from these theories. The FRONTS 87 experiment was designed to provide observations to test these hypotheses. We reiterate these here as our discussion of the dropsonde data will be centred around them.

(1) Ana and kata/split frontal structures are determined by the large-scale flow, in particular the structure of the potential vorticity (PV) field.

(2) Frontal motion is balanced on horizontal scales down to about 50 km.

(3) There is neutrality to slantwise moist ascent on the 100 km scale, but convective instability, on occasion, on scales less than 10 km.

(4) The periodic structure of line convection along the front is due to a dynamical instability.

(5) Precipitation loading and diabatic processes are a substantial influence upon mesoscale circulations at and near cold fronts.

(6) The surface cold front behaves like the head of a density current on a scale of a few kilometres.

(7) Frontal waves are associated with either upper or lower tropospheric potential vorticity anomalies.

(8) Most of the ageostrophic convergence in the frontal region is in the boundary layer.

This paper will be primarily concerned with mesoscale issues and the following is a discussion of three key dynamical ideas which recur in these hypotheses.

(a) Breakdown of balanced motion

The semi-geostrophic (SG) model of frontogenesis is the main theoretical framework in which to describe frontal dynamics. The review by Hoskins (1982) provides an introduction to this model. In essence the front is treated as a quasi-two-dimensional flow in which there is a close degree of geostrophic balance in the along-front direction. To maintain the implied thermal wind balance of that flow, if the synoptic-scale flow is forcing frontogenesis or frontolysis, there is an ageostrophic cross-frontal circulation. This was described by Sawyer (1956) and Eliassen (1962) and the circulation is now referred to as being described by the Sawyer–Eliassen equation. The degree of along-front thermal wind balance is related to the accuracy of the semi-geostrophic model.
However that model does not require that the imbalance be exactly zero; rather that the acceleration of the secondary circulation be small compared to the accelerations across the front due to the Coriolis and pressure-gradient forces. In fact the SG model can predict the thermal-wind imbalance (see Keyser and Reeder 1988).

This local breakdown of thermal wind balance, if the acceleration of the cross-frontal circulation becomes large, is manifested in inertia-gravity waves just above the ground in the core of the frontal ascent. Orlanski and Ross (1984) suggested that the ageostrophic motion associated with such waves might act to halt frontogenesis. In simulations using primitive equations, such as described by Garner (1989), it appears that this onset of wave activity plays an insignificant role in frontal dynamics; the frontogenesis proceeds without substantial modification. The other, and more potent, way in which the frontal balance can be perturbed is via the onset of secondary instabilities such as Kelvin-Helmholtz, convective, or symmetric instability. As the vertical shear increases because of frontogenesis the possibility exists that the Richardson number, based on the alongfront shear, will fall below 1/4. This will result in K-H billows aligned across the front; it can be shown that this may occur if the vertical component of the absolute vorticity exceeds a value of about 4f. Furthermore, as is shown by Thorpe et al. (1987), the horizontal scale of these billows across the front scales as the horizontal temperature difference across the front; for a difference of 5 K the scale is about 50 km. This means that for such a front we would expect to see regions with the Richardson number less than 1/4 and with horizontal scale greater than or of the order of 50 km, only if the vorticity is larger than about 4f. This prediction will be considered in the light of the FRONTS 87 observations.

It can be seen that the SG model goes a long way towards predicting details of the circumstances under which the model becomes inaccurate. The aim here is to diagnose from the data these relatively simple indicators of geostrophic breakdown. Convective and symmetric instability are other ways in which such a breakdown can occur; these are considered later. First we need to summarize the key role that potential vorticity ideas play, on the mesoscale, at fronts.

\( (b) \) Potential vorticity on the mesoscale

The potential vorticity (PV) is generally accepted as being a key dynamical variable for synoptic-scale flow (see the recent review by Hoskins et al. (1985)). On that scale it can be assumed that PV is approximately conserved, and further, owing to the degree of geostrophic balance, it can be inverted to find the flow and temperature distribution. The former property allows air-parcel trajectories to be deduced under the assumption of dry-adiabatic motion, whilst the latter property is generally referred to as the ‘invertibility principle’. The degree of conservation of PV is determined by timescales; for time-intervals greater than a few days diabatic processes such as radiation and latent heating will disrupt this conservation on the synoptic scale.

Whilst the use of PV ideas to describe synoptic-scale dynamics is becoming generally accepted, its use on the mesoscale is novel. Here the case for the use of PV ideas in describing the mesoscale dynamics of fronts will be presented. As discussed in the previous section, geostrophic balance at fronts is expected to apply down to horizontal scales of about 50 km. Hence it is likely that the invertibility principle is applicable down to these scales. It should be noted that it can indeed be used at any scale to deduce a measure of the ageostrophy of the flow. In contrast to the invertibility principle the use of PV as a conserved variable, and thus as a tracer, is definitely not possible on the mesoscale at fronts, where latent heat release is known to play a dominant role. In fact
with these data we will show that the $PV$ distribution is specifically valuable as an indicator of the effects of diabatic processes.

The saturated nature of at least the ascent in a frontal zone suggests the use of equivalent $PV$ and equivalent potential temperature; both of these are conserved in saturated adiabatic flow. Of course the descent in a front may be unsaturated, so we might expect that in different regions of the flow different conserved variables have to be considered. This is a complication, but one, we shall argue, that does not detract from the efficacy of the potential vorticity view of the mesoscale dynamics of fronts. To pursue the uses of $PV$ thinking we need to recall the role it plays in the SG theory of frontogenesis. The 'classical' models of cyclogenesis and particularly frontogenesis assume dry-adiabatic dynamics and, further, a uniform $PV$ distribution (see Hoskins and Bretherton 1972). This uniformity places a real constraint on the frontal dynamics and leads to the important feature of an intense front with large vorticity and variations in static stability though with no variations in $PV$. The constraint being spoken of here can be interpreted most simply in geometrical terms but with the important new facet of saturated ascent in the frontal zone. The following description is based on the SG models of moist frontogenesis presented by Thorpe and Emanuel (1985), Emanuel et al. (1987), and Joly and Thorpe (1989).

Consider, by analogy with the dry classical model mentioned above, a frontal zone with uniform equivalent $PV$ defined as

$$PV_c = \frac{1}{\rho} \zeta \nabla \theta_c.$$  \hspace{1cm} (1)

There is some evidence from the work of Emanuel (1988) and others that frontal zones are characterized by uniform $PV_c$; this is a hypothesis which we will be assessing in this paper. Furthermore the uniform value is thought to be near zero. Hence either the vorticity vector $\zeta$ and the gradient of $\theta_c$ are both small or else they are aligned approximately at right angles to one another. (Note that the figures presented in this paper have a horizontal scale which is compressed relative to the vertical scale by a factor of the order of 100. Therefore caution must be exercised when assessing the relative orientation of variable surfaces and vectors.) Hence in the lower troposphere at a front the vorticity vector points in the direction nearly parallel to the moist isentropes. The constraint of uniform $PV_c$ dictates two facets of a front which we shall show are commonplace: the near-zero absolute vorticity in the upper troposphere and the extremely large vorticity near the surface front.

Furthermore these models can be used to deduce the (dry) $PV$ distribution at the front; this being required for the invertibility notion alluded to earlier. It is clear that the $PV$ will be near zero in the upper troposphere with a large positive value in the lower troposphere. These anomalies in $PV$ are due to the latent heat release implied by the saturated frontal zone. It is plausible to suppose that the latent heat release maximum will be in the lower to mid troposphere thus implying an increase in $PV$ in the lower front and a decrease in the upper front.

This description can be formalized by recalling the equation for the rate of change of $PV$:

$$\frac{D(PV)}{Dt} = \frac{1}{\rho} \nabla \cdot (\theta \nabla \times F + \zeta \frac{D\theta}{Dt}).$$  \hspace{1cm} (2)

This equation shows that frictional forces $F$ and diabatic processes lead to a rate of change of $PV$ following air parcels. In particular, latent heat release in a cloud layer will
increase $PV$ 'below' the heating maximum and decrease the $PV$ 'above' the maximum, in the sense that these changes are oriented along the vorticity vector and hence along the $m$-surfaces. Note that the divergence form of this equation means that these local increases and decreases cannot change the mass-weighted average $PV$ unless the friction and diabatic processes occur at the ground.

What then are the essential reasons for the use of $PV$ ideas in a partly saturated front? As just presented, these can be summarized as follows:

**Equivalent potential vorticity**
- it is conserved in either two-dimensional or saturated inviscid motion (see next section);
- it is hypothesized to be approximately uniform (zero?) in frontal zones;
- regions of negative values indicate the possibility of conditional symmetric instability or convective instability (see next section);
- changes to $PV_z$ are produced in unsaturated three-dimensional inviscid motion or as a result of radiation, melting, and surface fluxes of moist entropy (see next section).

**Potential vorticity**
- it is conserved in unsaturated inviscid motion;
- anomalies indicate either advection or the dominance of latent heat release/uptake and friction (the lower frontal positive anomaly being a good indicator of the frontal position);
- the distribution of $PV$, along with the $\theta$-distribution on suitable boundaries, can be used to obtain the geostrophic flow from the invertibility principle (then the ageostrophic flow can be estimated from the mesoscale observations of the wind field);
- local extrema in $PV$ may be implicated in secondary barotropic/baroclinic instabilities (frontal waves) as discussed by Joly and Thorpe (1990).

One point in each of these lists relates to the possibility of secondary instabilities, a feature which will be a recurrent theme of the discussion here. In particular we now consider the notion of conditional symmetric instability.

(c) **Conditional symmetric instability**

Another key insight given by the role of $PV$ in the SG theory of (dry) frontogenesis is that the Sawyer–Eliassen equation for the cross-frontal circulation is elliptic if $PV > 0$ but becomes hyperbolic if $PV < 0$. In physical terms this means that an instability must be present if $PV \approx 0$, and this is referred to as symmetric instability. The name of the instability arises since the flow, and the instability, is two-dimensional by hypothesis. An illuminating way to write the $PV$ for this two-dimensional flow is by defining $m = v + fx$ giving

$$PV = \frac{\partial m}{\partial x} \frac{\partial \theta}{\partial z} - \frac{\partial m}{\partial z} \frac{\partial \theta}{\partial x} = J(m, \theta).$$  \hspace{1cm} (3)

Here the density has been assumed to be $1$ kg m$^{-3}$ for convenience. The variable $m$ is the absolute angular momentum on an $f$-plane, normalized by the distance of that plane from the axis of rotation. It is often referred to as the absolute momentum and whilst this is an inaccurate nomenclature we follow it here for consistency. Note that for a dry inviscid two-dimensional flow $PV$, $m$ and $\theta$ are conserved. Thus Eq. (3) shows that the
$PV$ is proportional to the area between the $m$ and $\theta$ surfaces. Also the instability relies on the relative slopes of these two surfaces; $m$ shallower than $\theta$ being unstable.

It is supposed that the criterion for symmetry instability, $PV \ll 0$, is rarely satisfied in the atmosphere but that for conditional symmetric instability (CSI), $PV_c \ll 0$, has been proposed to account for frontal rainband formation (see Bennetts and Hoskins 1979). The relative orientation of the $m$ and $\theta_c$ surfaces therefore plays a crucial role in CSI. In a frontal zone both these surfaces are quasi-horizontal, and if the $\theta_c$-surface is steeper than the $m$-surface then the criterion for CSI, $PV_c \ll 0$, is satisfied. Attention is focused by Bennetts and Hoskins (1979) on the descent in the CSI circulation being unsaturated. However saturated descent, which at times appears a better model from our observations, does not alter the stability criterion.

The parcel model of the instability, as described by Thorpe et al. (1989), shows that in the cross-frontal plane there is a wedge of instability which is bisected by the moist isentrope and includes the $m$-surface running through the point of parcel displacement. Further it is known from nonlinear simulations, such as described by Thorpe and Rotunno (1989), that the parcel motion is oriented approximately along the $\theta_c$-surfaces; consequently parcels cross the initial orientation of the $m$-surfaces, leading to a characteristic buckling of those surfaces. The reader is referred to Fig. 2 (b) of Thorpe and Rotunno (1989) for an example of the buckling process. The parcel model gives a clear physical picture of the instability. A displaced parcel suffers a horizontal force, $fm'$, towards its 'home' $m$-surface and a vertical force, $g(\theta'/\theta_0)$, towards its 'home' $\theta_c$-surface. Here a prime indicates the difference between the parcel value and the value of its current surroundings. If the resultant of these forces takes the parcel further from its point of displacement then the flow is unstable to such slantwise displacements. Most parcel kinetic energy can be derived by a displacement close to the $\theta_c$-surfaces; thus arises the apt description that CSI is inertial instability on moist-isentropic surfaces.

There are some key factors to bear in mind when analysing data with a view to deducing the presence, or not, of CSI.

- $m$ is only conserved in a two-dimensional flow;
- the criterion for CSI is only formulated for a geostrophic two-dimensional flow;
- primary observed quantities can be used to assess the susceptibility to CSI by examining the relative slopes of $m$ and $\theta_c$ surfaces; the estimation of $PV_c$ involves possibly inaccurate derived quantities;
- data of high vertical and horizontal resolution are required;
- these criteria are also satisfied, trivially, if the atmosphere is susceptible to convective instability; CSI refers only to convectively stable situations.

We conclude this section by recalling the equation for the rate of change of $PV_c$ first quoted by Bennetts and Hoskins (1979):

$$\frac{D(PV_c)}{Dt} = \frac{1}{\rho} \nabla \cdot \left( \theta_c \nabla \times \mathbf{F} + \frac{\partial}{\partial \theta} \frac{\nabla \theta_c \times \nabla \theta \cdot \mathbf{k}}{\rho \theta_0} \right) + \frac{g}{\rho \theta_0} \nabla \theta_c \times \nabla \theta \cdot \mathbf{k}. \quad (4)$$

Thus for a saturated or two-dimensional inviscid flow $PV_c$ is conserved. The last term represents a production or destruction of $PV_c$ in a three-dimensional sub-saturated flow. The other terms are due to friction or sources/sinks of $\theta_c$ such as radiation, melting or surface fluxes. Given that the criterion for CSI is in terms of $PV_c$, then it is important to analyse the production of such a condition from these terms. It is clear, in doing this, that it is necessary to quantify three-dimensional effects; this will not be considered directly in this paper.
3. SUMMARY OF THE INTENSIVE OBSERVING PERIODS

In the paper by Machin (1988) all of the eight intensive observing periods (IOP) of the FRONTS 87 programme are summarized. In Table 1 we present a list of the eight events. As can be seen no dropsonde data were obtained for IOP 5 and 6, so in this paper there are six cases to be considered. Of these six cases IOP 1 and 4, being quasi-stationary and an occlusion respectively, did not come directly under the definition of ideal systems laid down before the experiment, but they did give interesting data. Thus we are left with IOP 2, 3, 7, and 8 as the four cold fronts with good data coverage from the C-130 and all other data sources. In this paper we will therefore concentrate on these four examples.

**TABLE 1. SUMMARY OF THE INTENSIVE OBSERVING PERIODS**

<table>
<thead>
<tr>
<th>IOP</th>
<th>Start</th>
<th>End</th>
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</thead>
<tbody>
<tr>
<td>1</td>
<td>1200 18/10/87</td>
<td>0000 20/10/87</td>
<td>quasi-stationary front; frontal waves</td>
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<td>2</td>
<td>0300 11/11/87</td>
<td>1500 12/11/87</td>
<td>multiple front; rainbands</td>
</tr>
<tr>
<td>3</td>
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<td>quasi-two-dimensional rainbands</td>
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<tr>
<td>4</td>
<td>0000 13/12/87</td>
<td>1200 14/12/87</td>
<td>occlusion moving SW to NE</td>
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<tr>
<td>5</td>
<td>1200 17/12/87</td>
<td></td>
<td>aborted</td>
</tr>
<tr>
<td>6</td>
<td>0000 05/01/88</td>
<td>1200 06/01/88</td>
<td>aircraft radar malfunction; no data</td>
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<tr>
<td>7</td>
<td>0000 09/01/88</td>
<td>1200 10/01/88</td>
<td>no dropsonde data; air traffic problems</td>
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<td>8</td>
<td>0600 12/01/88</td>
<td>1800 13/01/88</td>
<td>quasi-two-dimensional weak line convection</td>
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**TABLE 2. START-TIMES AND END-TIMES (GMT) OF RUNS**

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<tr>
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In Fig. 1 a summary of the synoptic charts for each of these fronts is shown including the North Atlantic sector mean-sea-level pressure, 500 mb, and 250 mb height charts, and the 6-hourly surface frontal progression with the location of the soundings. (In Table 2 the times of the start and the end of each run are given.)

The pre-experiment flight plan was in fact rather closely adhered to in all cases. In IOP 7 and 8 it was decided not to fly a fifth run but rather to extend the length of some or all of the other four runs in order to be able to describe both the upper and lower frontal surfaces more fully. Few sondes failed, but when this happened another one was launched immediately; over 90% of the maximum possible retrievable data were retrieved from the sondes.

The following is a summary of the weather characteristics of these events as noted by Machin (1988), which does not attempt to provide a full log of each case but rather to give the general character of the fronts as perceived at the time of the experiment.
Figure 1. Manually-drawn Meteorological Office analyses of the mean-sea-level pressure, 500 mb, and 250 mb height for the four main events from FRONTS 87, IOP 2, 3, 7, and 8. The contour intervals are 4 mb, 6 dam, and 12 dam respectively. Also given is the 6-hourly progression of the surface frontal position with the dropsonde pattern superimposed. Start-times and end-times of each run are given in Table 2.
Figure 1. Continued.
IOP 2: As the cold frontal zone crossed the British Isles more than one surface discontinuity was analysed. At 12 GMT the routine radar network showed line convection within a band of precipitation 150 km wide. Frontal speed was estimated at 17 m s\(^{-1}\) (but somewhat smaller over the dropsonde area) with a marked decrease of wind speed after frontal passage, and drops of temperature and dew-point of less than 2 degC. There were precipitation bands ahead of the cold front. It was felt at the time to be a 'classical', or rearward-sloping, or ana-cold front. The second discontinuity consisted mostly of a dew-point fall and cloud clearance. As the front approached Brest around 1930 GMT a minor propagating wave temporarily slowed its progress.

IOP 3: After the progression of a frontal wave, which was over central England at about 12 GMT, the front was thought to be a very good example of a rearward-sloping cold front. Extensive moderate and locally heavy precipitation was observed from several broad rainbands. The frontal passage was accompanied by a wind veer greater than 60 degrees, locally reaching 120 degrees, and a temperature drop of up to 4 degC. Upper winds over the British Isles were in excess of 75 m s\(^{-1}\). A broad post-frontal rainband was evident over England and Wales. The frontal intensity was diminishing markedly while passing through the dropsonde area.

IOP 7: The front was essentially two-dimensional but with pulses of enhanced rainfall (flat waves) moving along the front. At the surface front wind veers of about 20 degrees were observed with a temperature fall of about 2 degC. The parent low had deepened explosively during the latter half of 8 January.

IOP 8: This front exhibited waves which moved quickly northwards along its length. Rainbands were evident with strong line convection at Brest. A low-level jet speed of 33 m s\(^{-1}\) was reported from the Camborne ascent at 21 GMT on 12 January. The approximate frontal speed was 10 m s\(^{-1}\) from a direction of 280 degrees. A marked wind and temperature change accompanied the frontal passage.

4. FRONTAL STRUCTURE FROM THE DROPSONDE DATA

In this section various common structures revealed by the dropsonde data are presented. The subsections herein do not necessarily relate to the individual hypotheses quoted in section 2. In the concluding section of the paper we return to those hypotheses so as to unify the structures described here into a coherent dynamical picture of active cold fronts.

(a) Characteristics and analysis of the dropsonde data

The Meteorological Office dropsonde used was essentially that described by Ryder et al. (1983), with newly-designed transmitters and LORAN-C receivers. In order to operate over the large area of the FRONTS 87 experiment the aircraft processing system and algorithms were redesigned to use cross-chain LORAN operation for windfinding. Use of additional signals from the French LORAN-C transmitters considerably improved
and extended the wind-finding capability compared to that described by Ryder et al. The wind-finding algorithms were based upon the least-squares method of Passi (1974) using combinations of all adequate signals weighted according to their signal-to-noise ratio (Clough et al. 1987). Comparisons with radar-tracked radiosondes at Bracknell showed one-minute-averaged winds of 1 m s\(^{-1}\) or better accuracy, and estimated errors throughout the experimental area were consistent with this value. Humidity measurements were carried out using the VIZ carbon hygistor, which has a typical accuracy of about 5% above 30% relative humidity, although it has little response below 20%. Response times are highly temperature-dependent, but generally quite short compared to most radiosonde humidity sensors. For example, Nash and Schmidlin (1987) estimate from trial soundings time constants of order 1 s at 0°C and 20 s near -30°C. The main departure from this performance is due to a lag of up to 30 mb on leaving deep layers of wet, icing cloud. Sensor lag is most evident where excessive humidity gradients occur above 5–6 km as the sondes adjust to moist warm-sector air from dry aircraft conditions at 7–8 km. For various reasons the pressure measurements were frequently found to be in error by 2 mb or more at the sea surface so that evaluation of accurate geostrophic winds was not possible; instead deviations from thermal wind balance were used for this study.

The observations were analysed using either a Cressman filter approach or a two-dimensional smoothing spline algorithm (for details see Thiebaux and Pedder (1987)). These two methods proved to give comparable results and the analyses shown here are from the spline method using a fit to the data with r.m.s. residuals of 0.3 m s\(^{-1}\), 0.5 K, 5%, and 0 mb. Subsequent manipulations of the analysed fields were based on finite differencing over scales which were considered to be comparable with the smallest unambiguously resolved scale represented by the observations, namely 20 km in the horizontal and 500 m in the vertical.

(b) Conserved variables \(m\) and \(\theta_e\)

A dynamically relevant analysis can be made by examining the structure of variables which are conserved under certain circumstances. The thermodynamic variable \(\theta_e\) (or \(\theta_w\), the wet bulb potential temperature) is conserved, in the absence of surface fluxes or radiative effects, by unsaturated or saturated motion. We choose here to combine analysis of \(\theta_e\) with that of the 'absolute' momentum, \(m\). The latter obeys the exact equation:

\[
\frac{Dm}{Dt} = f u_g + F_v
\]  

where \(u_g\) is the geostrophic component of the cross-frontal flow and \(F_v\) is the \(y\)-component of the frictional force. So in the absence of both friction and an along-front temperature gradient the absolute momentum is conserved in a frame moving with the front. Away from the surface and in a two-dimensional front the sources/sinks of \(m\) will therefore be zero.

In Fig. 2 cross-sections of \(m\) and \(\theta_e\) are shown. The structure of the \(\theta_e\)-field was the most reproducible feature of all the fronts observed in the experiment. An extremely sharp horizontal gradient is evident, sloping with height, separating two airmasses of relatively homogeneous character. Below about 850 mb this gradient reverses slope, leading to a 'nose' structure in \(\theta_e\). The surface contrast in \(\theta_e\) is typically 200–300 km behind that at 850 mb. This feature is likely to be due to the strong influence of sea-surface fluxes of moist entropy as the front progresses across the North Atlantic.
Figure 2. Vertical cross-sections from the dropsonde data of the distributions of $\theta_s$ and $m$ in run 3 from IOP 2, 3, 7 and 8. The ordinate is height and the abscissa is the horizontal distance in the across-front direction. The tick-marks on the ordinate are every 1 km, the tick-marks on the abscissa are every 50 km and those on the upper boundary show the release location of the sondes used in the analyses. Note that the horizontal scale representing 50 km is different for each run because of different total lengths. Note also that the absolute momentum involves an arbitrary origin constant so that the absolute value has no significance. The sonde release location is discretized to 10 km giving a slight, and unimportant, displacement of the upper tick-marks.
Another feature common to all cases is the occurrence of folds or steps in $\theta_e$, where the frontal surface intersects the main rear cloud edge, above or within the melting layer. This is particularly pronounced in IOP 3 run 3 at a height of between 3 and 5 km, and is evidence of middle-level instability. Similar features are also evident in diagrams of Browning and Harrold (1970) and a schematic from Matejka et al. (1980), though without detailed documentation or discussion. In the cold fronts observed here these features usually coincided with the level of the strongest forward cross-frontal flow, and are attributed, at least in part, to the evaporation and, possibly, melting of snow: they are discussed more fully in a later section.

The absolute momentum, $m$, is another conserved variable, in the absence of friction. Its local slope in the $x-z$ plane is $dz/dx = -\xi/\nu_z$ so that in a barotropic flow the contours of $m$ are vertical, whereas in the baroclinic flow at a front they develop a substantial slope. For example if $\nu_z \approx 10 \text{ m s}^{-1}$ per km and $\xi = f$ then the slope $dz/dx \approx 1$ in 100. In the SG models of frontogenesis, $m/f$ is used as a horizontal coordinate and has a smooth slope of this order. The observed $m$-surface distributions in Fig. 2 clearly slope as just described in the frontal zone but with a ‘superimposed’ mesoscale structure. We identify three such structures:

- ‘buckled’ $m$-surfaces on the warm side of the frontal zone but rearward of the surface front location;
- a region in the warm air in the lower troposphere of uniform $m$ structure (particularly evident in IOP 8);
- in the frontal zone particularly in the mid to upper troposphere the $m$ and $\theta_e$ surfaces have a similar slope.

The buckled $m$-surfaces in IOP 7, consistent with negative absolute vorticity, are reminiscent of those found in the simulations of CSI described earlier. This is particularly so when we note that the $\theta_e$-surfaces in this region are not buckled and indeed clearly indicate appreciable moist stability. The horizontal scale of the buckled surfaces is of the order of 100 km. Referring back to hypothesis 3 in section 2 we see from these data that neutral conditions apply in the mid to upper frontal zone but that locally there are, possibly, unstable slantwise displacements. It now becomes a key issue to determine whether CSI is the only cause of such buckled $m$-surfaces. If it is, then this is the first evidence in the literature of active CSI. Other mechanisms may be able to distort the $m$-surfaces in this way; for example friction and three-dimensional effects. In fact there is a suggestion by Browning and Papad (1973) of the existence of a similar structure from the earlier radar observations of the Scillonia project in which there was buckling in the along-front wind field pattern, which almost certainly corresponded to buckled $m$-surfaces, although absolute momentum analyses were not carried out at that time.

An important recent observational result on frontal rainbands is given by Emanuel (1988) in which soundings were attempted by flying an aircraft along an $m$-surface. The results as given there suggest that the temperature profile is essentially moist-adiabatic along the several $m$-surfaces that were so traversed. This corresponds to $PV_e \approx 0$, which is the condition for neutrality to CSI. The important questions from that study are: how representative of the front are the few soundings reported by Emanuel (1988), and does the neutral condition arise from previous unstable conditions? We here find that the near-neutral condition is representative of the mid to upper frontal zone and that there is evidence of active CSI.

In Figs. 2 and 3 the homogeneous zone of $m$ in IOP 8 is identified in all runs in the warm air and is clearly two-dimensional. Such a zone requires a region with near-zero
Figure 3. Vertical cross-sections from IOP 8 showing the $m$-distributions on runs 1, 2, and 4. Together with that for run 3 given in Fig. 2 these show the full along-front variation of absolute momentum. Each run is approximately 100 km displaced along the front compared to the next, see Fig. 1. Note that the horizontal scale representing 50 km is different for each run because of different total lengths. The region of relatively well-mixed absolute momentum in the warm air ahead of the front is apparent in every run.
absolute vorticity and along-front shear. The region at low levels between the frontal zone and the low-level jet is one where the vertical shear is often small. Above this jet the shear is negative so the m-surfaces must slope in the opposite sense to those in the frontal zone. Hence the homogeneous zone is located in the region where the m-surfaces spread out as they emerge from the frontal boundary layer. However, pure advection of the m-surfaces will conserve PV and thus be incapable of producing a ‘lens’ of zero PV. Therefore the kinematical description of this zone based on advection by the cross-frontal flow is probably incapable of accounting for the development of this structure, and possible dynamical mechanisms for its formation must be explored.

A proposal for the development of such regions has been made by Holt and Thorpe (1991) in which air moving rapidly upwards out of the boundary layer in the frontal line convection inflates a ‘lens’ of uniform m-values. The key to this process is that this air on leaving the line convection has anomalous inertia, m'. Under the action of the inertia force, fm', the air executes a slantwise adjustment along moist isentropic surfaces to return to its ‘home’ m-surface. As air is continually fed upwards out of the boundary layer this ‘lens’ is inflated to reach a steady-state size dictated by the magnitude of the m-anomaly. An important aspect of this mechanism, however, is that it requires conditions of weak stability to CSI.

Another possible mechanism involves the production of a region of uniform m by the repeated collapse or mixing of buckled m-surfaces generated by active CSI. As the buckled m-surfaces represent a local anomaly in inertia they will be liable to irreversible mixing processes, which would ultimately lead to the generation of zero PV_e. This is the process that is implicit in Emanuel’s idea of fronts being characterized by neutral conditions. However as noted by Thorpe and Rotunno (1989) there is a problem in supposing that mixing will produce a region of zero PV_e as a result of CSI circulations occurring in a flow with PV_e ≤ 0 locally. As can be seen from Eq. (4) there is no requirement for the traditional down-gradient mixing of heat and momentum to produce a down-gradient mixing of PV_e; the latter would be necessary to account for the condition of neutrality, or PV_e = 0. This remains an unsolved problem in CSI dynamics and is clearly of some importance in understanding the structures that are observed at fronts.

(c) ‘Classical’ structure of along-front flow and θ

The traditional method of presenting frontal structure is in terms of a cross-frontal cross-section of along-front flow (here denoted by v) and potential temperature, θ. As emphasized above, the more dynamically relevant analysis is in terms of conserved variables, but we here digress to pick out two features of interest in the along-front flow and θ-fields.

In Fig. 4 we show this analysis for run 3 of IOP 8, which was the longest highly resolved run of the experiment. The upper-level and low-level jets are evident as is the dramatic windspeed reduction in the cold air just behind the front. The apparent connection between the upper and lower jets is of note, as is its very short horizontal scale. The thermal structure is indicative of a sloping baroclinic zone but is notable in its lack of substantial horizontal gradients in the lower troposphere. In this case it is clear that the warm air has a significantly larger static stability than the cold air. The dynamical role of such a stability difference between the two ‘airmasses’ is discussed by Thorpe (1990), and for the difference observed here it is noted that upper frontogenesis is likely to be accentuated. Such stability differences between the airmasses are due, for example, to surface heat and moisture fluxes.

The difference in intensity of gradients between the θ and θ_e fields is evidence of the importance of humidity gradients in such cold fronts. From the definition of θ_e it can
be shown that the fractional change in $\theta_e$ across the front at any given pressure level is related to the temperature change and the relative humidity change via the formula:

$$\frac{\delta \theta_e}{\theta_e} = \left(1 + \frac{L^2 q}{R_v c_p T_{sat}}\right) \frac{\delta T}{T} + \frac{Lq}{c_p T_{sat}} \frac{\delta RH}{RH}$$

(6)

where $RH$ is the relative humidity, $T_{sat}$ is the temperature at the level of saturation of a lifted parcel, and $q$ is the specific humidity. For typical values in the lower troposphere at around 700 mb of $T = 270$ K and $q = 5$ g kg$^{-1}$ and assuming that $T_{sat} \approx T$ then the percentage change becomes

$$\frac{\delta \theta_e}{\theta_e} \approx (100 + 86) \frac{\delta T}{T} + 4.5 \frac{\delta RH}{RH} \%.$$  

(7)

So for a temperature change across the front at low-levels of 5 K and a humidity change from 100% to 50% we find that $\delta \theta_e/\theta_e \approx 3.3 + 2.3\%$.  

Figure 4. Vertical cross-sections from run 3 of IOP 8 of the distributions of along-front flow. Contour interval 5 m s$^{-1}$; $\theta$, contour interval 2 K.
From these illustrative figures we see that $\delta \theta_c / \delta T = 3$. If the front had had no humidity gradient then $\delta \theta_c / \theta_c \approx 3.3\%$ and $\delta \theta_c / \delta T = 2$ whilst if it had been dry then $\delta \theta_c / \theta_c \approx 1.8\%$, and $\delta \theta_c / \delta T = 1$. In the upper troposphere the humidity content is so low that the frontal transition is entirely dominated by the change in temperature. Clearly the humidity contrast and the temperature contrast contribute a similar amount to the change in $\theta_c$ across the front, in the lower and mid troposphere.

A feature which is common in these data is the tendency for alignment between the contours of $\nu$ and $\theta$ in the mid to upper troposphere within the sloping frontal surface. That this might be the case was argued by Gill (1982) on the basis of the Hoskins and Bretherton (1972) frontal model. Consider a situation in which potential vorticity is conserved and is relatively uniform before the front forms. As the developed front is a region of large horizontal and vertical gradients then, for the $PV$ to remain constant, the terms in the expression for $PV$ involving the relative vertical vorticity and the horizontal vorticity must nearly cancel, i.e.

$$\frac{\partial \nu}{\partial x} \frac{\partial \theta}{\partial z} - \frac{\partial \nu}{\partial z} \frac{\partial \theta}{\partial x} = J(\nu, \theta) = 0. \quad (8)$$

Therefore the remaining term in $PV, f(\partial \theta / \partial z)$, approximately equals the pre-frontal tropospheric $PV$. For the above Jacobian to be zero implies that the $\nu$ and $\theta$ contours coincide. This argument is only plausible for an air-flow in which the $PV$ is conserved as the front develops. As discussed previously in this paper, diabatic processes change the $PV$ locally, so it is unlikely that $PV$ is in fact conserved. We will return to this later when we discuss the $PV$ structure in detail. If $PV$ does change then there is no requirement for the Jacobian in Eq. (8) to be zero. It is plausible however that the changes in the $PV$ in the mid or upper troposphere are smaller than those of the individual gradients that make up the $PV$. This must be the case to explain the close parallelism of the two surfaces.

\section*{(4) Thermal wind imbalance (TWI)}

The degree of geostrophic balance is one of the key issues in the assessment of the direct validity of the semi-geostrophic description of frontogenesis. The prediction is that any zone that is substantially ageostrophic in the along-front direction will be of horizontal scale less than or of the order of 50 km. It is unfortunate that the C-130 did not have available a high-altitude radar altimeter which would have allowed a surface pressure analysis of resolution comparable to that of the sounding data. As a consequence we do not have a direct way of obtaining the geostrophic or the ageostrophic wind fields. Later papers will provide estimates of these by applying objective analysis techniques to these and other data sources. Here we present only the thermal wind imbalance, which can be accurately assessed from the dropsonde data. It is defined as

$$TWI = \frac{\partial \nu}{\partial z} - \frac{g}{f \theta_0} \frac{\partial \theta}{\partial x}. \quad (9)$$

For the high-resolution run from IOP 7 we show in Fig. 5 the $TWI$ field. The error in computing $TWI$ from the observations is about 0.003 s$^{-1}$, which is large but as we now show there are coherent patterns of $TWI$ with values significantly above this noise level. From Fig. 5 there are positive values of $TWI$ of order 0.01 s$^{-1}$ in the lowest kilometre, and negative values aloft with peak values in excess of 0.01 s$^{-1}$. This observation does not agree with the conclusion of Browning and Pardoe (1973) that the low-level jet is in thermal wind balance. However, the large values of $TWI$ are confined to the frontal
zone, with a scale of about 50 km in the direction at right angles to the sloping frontal surface. Because of that slope, of course, the total horizontal extent of the region is several hundred kilometres. The sub-geostrophic nature of the TWI away from the boundary layer is significant because it is, for example, what one would expect within a CSI roll circulation. For a two-dimensional flow it is easy to show that

$$\text{TWI} = \frac{1}{f} \frac{\partial}{\partial t} \frac{\partial\psi}{\partial z}$$

So the TWI is proportional to the rate of change of the \( \psi \)-component of vorticity; this vorticity is \( \nabla^2 \psi \), where \( \psi \) is the streamfunction of the cross-frontal flow, in the absence of frontal forcing. (If there is frontogenesis then the vorticity will include an additional part due to the along-front temperature gradient). So negative TWI is consistent with either a growing symmetric instability roll with an anti-clockwise circulation, or an increase in time of a thermally direct cross-frontal circulation at a cold front.

The latter is exactly what is expected if a front is undergoing active frontogenesis or if there is diabatic forcing occurring on a timescale shorter than that of any geostrophic adjustment. The large magnitude of the TWI in the frontal zone is therefore indicative of the important role of latent heating in the dynamics. However it's limited horizontal extent implies that the semi-geostrophic model of frontogenesis is nevertheless still likely to be accurate.

(e) Humidity structure

As suggested in the last section, these fronts represent a substantial humidity contrast. A comparison of the cross-sections of \( \theta \) and \( \theta_e \) gives an indirect picture of this contrast, but here we show the humidity distributions in Fig. 6. A large variability is evident even within IOPs, because of frontal waves and the varying history of systems.

It is notable that the relative-humidity contours do not follow the frontal surface as characterized in \( \theta_e \). Thus although the cold air in the mid to lower troposphere is very dry, presumably because of substantial descent, the moisture transition at these levels is
Figure 6. Cross-sections of the relative humidity from run 2, IOP 2; run 3, IOP 3; run 3, IOP 7; and run 3, IOP 8. Contour interval is 15%.
100–300 km behind the surface front. This characteristic is likely to reflect both the effectiveness of convection in moistening these levels and the efficient evaporation of precipitation from the main frontal ice-cloud mass (Clough and Franks 1991).

Higher up, the frontal cloud mass does not extend along the sloping front into the upper troposphere, but terminates typically in mid troposphere with perhaps a narrow tongue extending further (particularly noticeable in IOP 7). The humidity transition tends to be vertically aligned or even sloping towards the warm air, with only occasional shallow intrusions of opposite slope evident in the highly-resolved runs. Possibly descent of the tropopause down the frontal zone prevents the rearward ascent of warm moist air to the upper front, but the factors determining the characteristics of the interface between upper and lower tropospheric air at fronts are not well known. The sloping structure may suggest simply that because of its greater latent heat content the warmer air ahead of the surface front has ascended progressively further until it is occluded. (Comparison with the \( \theta_c \) pattern for IOP 3, Fig. 2, is consistent with this view.) However it is possible that dynamical constraints imposed by the potential vorticity pattern under frontogenetic forcing may also influence the envelope shape.

\[ (f) \quad \theta_{es}\text{-field and tephigrams in the cold air} \]

An important aspect of the thermodynamics of moist air is the relationship of the vertical temperature profile to that of the saturated, or indeed dry, adiabat. A vertical sounding through a convective cloud will be saturated, with a moist adiabatic temperature profile. Alternatively if air has, at a previous time, undergone moist convection it may plausibly be adjusted to a condition in which its temperature structure is that of the moist adiabat; however it may now be subsaturated. Also along the sloping frontal zone the proximity of the contours of \( \theta_{es} \) to those of \( m \) will show an adjustment to neutral conditions to CSI. It is therefore illuminating to diagnose the saturated equivalent potential temperature, \( \theta_{es} \), structure at fronts. It should be remembered that \( \theta_{es} \) gives information only on the thermal structure and gives no information concerning humidity.

In Fig. 7 a cross-section of \( \theta_{es} \) is given for run 3 of IOP 7. Two features are of considerable interest. The first is the striking parallelism of \( \theta_{es} \) and \( m \) (given in Fig. 2) in the frontal zone from above the boundary layer to the uppermost of these observations. These surfaces are more closely parallel than those of \( \theta_c \) and \( m \). Therefore we might

![Figure 7. The distribution of \( \theta_{es} \) for run 3 of IOP 7.](image)
infer that the drier air of the upper frontal zone as well as the saturated air of the frontal cloud must have undergone an adjustment to this neutral state from conditions unstable to CSI. The result of this adjustment is the parallelism that is observed over virtually the whole depth of the sloping front. In a steady state front this must imply that the mechanisms attempting to destabilize the flow to CSI are acting on a similar timescale to those attempting to neutralize the atmosphere. The dynamics of these destabilization and neutralization processes to CSI are not well understood. It cannot be ruled out that the front attains this near-neutral state from pre-frontal conditions of stability but that the frontogenesis, in rotating the gradient of $\theta_w$ away from the vorticity vector, can produce conditions which are indistinguishable observationally from neutral whilst always maintaining stability to CSI.

![Diagram](image)

Figure 8. A tephigram from the westernmost dropsounding in the cold air from run 3 of IOP 7 showing the remarkably moist-adiabatic, but subsaturated, profiles.

The other feature of note is the vertical uniformity of $\theta_w$ above the boundary layer in the cold air behind the front. We emphasize this by showing in Fig. 8 a tephigram in this region. The air is clearly subsaturated but with a temperature profile, throughout the depth of the troposphere, lying almost exactly along a moist adiabat. This was an unexpected feature but one of some importance when it comes to the specification of characteristic reference profiles for such airstreams. Betts (1982) has described such profiles for tropical airstreams and notes there the proximity of the temperature profile to that of constant $\theta_w$; as these fronts are in the cooler mid latitudes the virtual effects are here negligible. In the warm air it is only the upper troposphere that exhibits such a uniform $\theta_w$ profile, but we note that owing to the location of the soundings the warm air is rarely sampled.
(g) Melting-layer structure

It is well known that as precipitation falls through the melting layer at 0 °C the absorption of the latent heat of fusion leads to a cooling of the air. Because of the finite timescale for melting as snow falls, this tends to produce a stable near-isothermal layer at 0 °C of significant depth (see Atlas et al. 1969 for a discussion of this process). Here we present in Fig. 9 a tephigram which shows this structure clearly. The melting layer has a depth of anything up to 500 m, which is comparable to the depth of the frontal zone itself. Note that immediately under this layer the temperature profile is dry adiabatic or even absolutely unstable. Also it appears that, despite any descent which might be associated with the cooling, the melting layer appears saturated. This condition is presumably maintained by the efficient evaporation of melting snow, as rain evaporation is incapable of maintaining saturated descent at vertical velocities much exceeding 1 cm s⁻¹ (Clough and Franks 1991).

![Figure 9](image)

Figure 9. A tephigram from a dropsonde in run 3, IOP 2 showing the structure of the melting layer. (The difference of the dew-point curve from precise saturation in the troposphere is due to a small calibration error in this particular humidity sensor.)

(h) Cross-frontal flow and its divergence

The cross-frontal flow is, in the SG model of frontogenesis, purely ageostrophic in the absence of along-front temperature gradients. For active frontogenesis it predicts a thermally direct cross-frontal circulation which would be consistent with cross-front flow towards the warm air at low levels and towards the cold air at upper levels. In Fig. 10 the cross-frontal flow, in the frame of reference moving with the front, is shown for one run in IOP 8. It is clearly consistent, in overall structure, with the thermally direct circulation envisaged in the SG model. However there are smaller-scale features not
predicted in the SG model. For example in the pre-frontal boundary layer there is strong flow towards the front, and just above the boundary layer a return flow is evident. The former feature is consistent with strong ageostrophic convergence in the boundary layer, which was asserted in hypothesis 8 of the experiment. Many of the other smaller-scale structures appear to be associated with moisture transitions.

Another important structure is shown in Fig. 10 from IOP 7. There is a shallow zone in the cold air under the frontal surface of enhanced flow towards the warm air. Comparison with Fig. 6 shows that this feature is confined to the cloud, towards its rear edge; this is also evident in run 4 which is not shown here. Such a feature is reminiscent of observations of flow beneath stratiform clouds in squall lines (see for example Smull and Houze 1987). In a front this is likely to be due to the evaporation of snow, which locally cools the air. Also a small drag is exerted on the air by the falling precipitation so the total effect is to enhance the descent in that region. As described by Eliassen
(1959) such a diabatic cooling produces descent along the sloping $m$-surfaces and so this cooling gives an enhancement of the cross-frontal flow. A discussion of the microphysics of frontal snow evaporation has recently been given by Clough and Franks (1991). Also they discuss the interaction of the consequent cooling with frontal dynamics. A key point from that paper is that the evaporation process is highly efficient for snow in the frontal environment where static stability restricts the vertical motion to being less than or of the order of $1 \text{ m s}^{-1}$. Hence the air is likely to be descending essentially moist-adiabatically in this region, and it is interesting to note from comparison with Fig. 2 that the forward current closely parallels surfaces of $\theta_e$, a pattern noted also in the highly-resolved fourth run. This theory cannot be fully tested here owing to the lack of an accurate direct measurement of vertical velocity on the mesoscale, but we take this zone of enhanced cross-frontal flow to provide support for their model. It should be noted that such enhanced flows were found near the lower rear part of the cloud canopy in all cases and that on occasion, as in this case, it is found within the frontal zone itself.

The clear evidence of evaporative cooling in these observations leads one to speculate on the density-current model of a cold front discussed in hypothesis 6 of FRONTS 87. The temperature deficit of the cold air is clearly being enhanced by diabatic processes, but at fronts there are important dynamical processes, such as the conservation of $m$ and synoptic forcing, not included in simple density-current models. Our observations indicate that geostrophic control exists almost down to the smallest resolved scales in the data, so it is unlikely that the density-current model of a cold front is applicable on these scales. Further discussion of these issues is given by Smith and Reeder (1988).

The kinematic computation of horizontal divergence, $D$, and vertical motion from serial radiosoundings has been discussed by Saarikivi and Puhakka (1989). Here we confine our analysis to $D = \partial u / \partial x$ as the other component $\partial v / \partial y$ is poorly observed with these data and is found, generally, to be small. Here the maximum expected error in $D$ is about $0.5 \times 10^{-4} \text{s}^{-1}$. The pattern of $D$ is clear and distinctive in the two best resolved cases, IOP 7 and 8, shown in Fig. 11. The patterns are composed of alternating sloping layers of convergence ($C$) and divergence, the layers lying nearly parallel to the $m$ or $\theta_e$ surfaces. On the basis of the simple moist-up, dry-down model of deformation-induced frontogenesis described by Thorpe and Emanuel (1985) we would indeed expect such layers to have a $D-C-D$ configuration. For these two cases a $D-C-D$ pattern is apparent, particularly in the lower troposphere; however what is totally unexpected, from a theoretical viewpoint, is the localization and intensity of the divergence. Values of the divergence are smaller than $-1 \times 10^{-3} \text{s}^{-1}$ over layers of width about 50 km. Browning (1983) refers to 'laminations of divergence' apparent in the radar observations of Browning and Harold (1970); this accords well with our observations, even though IOP 7 did not exhibit line-convection and so is a less intense front than that considered by those authors. As stated previously, we feel that in the locality of the frontal zone the descent is moist adiabatic and so the theoretical models of frontogenesis underestimate the divergence by neglecting evaporative cooling.

(i) Potential vorticity and related structures

Given the importance attached to the potential vorticity distribution on the mesoscale in sections 2 (b) and 2 (c) we here attempt to diagnose its structure from these data. The flight runs of the dropsonde deployment pattern lead naturally to a cross-frontal cross-sectional approach to the data analysis. Thus it is straightforward to estimate the $PV$ on the basis of Eq. (3) which assumes a two-dimensional flow. Attempts have been made to estimate the along-front derivatives by computing simple finite differences between
the runs, and this indicates that indeed these derivatives are small for many of the IOPs. However we prefer to postpone a fuller discussion of the three-dimensional aspects to a later paper in which a full three-dimensional objective analysis will be presented. Here we simply note that the PV and vorticity fields shown here are two-dimensional estimates of the full three-dimensional quantities. Clearly the calculation of this quadratic quantity is, in general, subject to considerable error; our estimate of this error for these mesoscale observations is about 0.1 to 0.2 PVU (potential vorticity units).

An insight into the factors making up the PV distribution is given by the geostrophic estimate of the PV. The potential vorticity can be approximated in the following way if the flow is in geostrophic balance:

\[
PV_g = \frac{\theta_0}{g} N^2 f \left( \frac{\zeta_g}{f} - \frac{1}{Ri_g} \right)
\]  

(11)

where \( N \) is the Brunt–Väisälä frequency, \( \zeta_g \) is the geostrophic absolute vorticity, and
\( Ri_g = N^2/(v_z g)^2 \) is the geostrophic estimate of the Richardson number in which \((v_z)g\) is the thermal wind shear. The suffix 'g' signifies that this is a geostrophic estimate of the full Ertel–Rossby PV for a two-dimensional flow. From this equation it is clear that the static stability, the Richardson number, and the absolute vorticity divided by \( f \) are the three parameters determining the potential vorticity. The last two of these are, of course, non-dimensional parameters. It is illuminating to look at these three factors separately as well as at the PV itself. Although the above form of PV requires the geostrophic wind we here show these parameters based on the wind itself.

In Fig. 12 we show some typical sections of \( N^2 \), absolute vorticity, and Richardson number \( Ri \). The frontal zone is characterized at low levels by a large vorticity maximum but also by a small stability anomaly. In contrast at mid and upper levels the front is clearly evident as a stable layer but has a weaker vorticity maximum. Also the vertical shear, or alternatively the horizontal temperature gradient, is large in the frontal zone (not shown here but see Fig. 2). Notice that the Richardson number, which is the ratio of two quantities which both have large maxima in the frontal zone, has a substantial minimum in the frontal zone. This can be understood from the following argument. It is easy to show that

\[
Ri_g = f^2 \frac{\bar{\theta}}{\Delta x} \frac{g\Delta z}{\partial \theta/\partial x}
\]

where \( \Delta x/\Delta z \) is the slope of the \( \theta \)-surfaces. For typical values \( Ri_g \approx 1.5/\bar{\Delta} \), where \( \bar{\Delta} \) is the horizontal temperature change per 100 km. Hence, given a constant frontal slope as the front develops, the Richardson number should decrease as the horizontal thermal gradient increases. (For \( Ri_g \approx 0.25 \) we require that \( \bar{\Delta} \) be about 6 K per 100 km.) This behaviour can be explained more rigorously from the semi-geostrophic model so that, at least in the mid troposphere, it can be related to the dynamics of frontogenesis. As the front collapses the horizontal and vertical gradients of \( \theta \) increase, but the cross-frontal circulation acts to diminish the vertical stability as warm air rises and cold air sinks. Thus there is a proportionately larger increase in the horizontal compared to the vertical gradient. Even at the low horizontal resolution of these data the minimum \( Ri \) is as low as 0.5. A key point as far as the validity of the SG model is concerned is that this zone of low Richardson number is confined to a zone of order 50 km in the horizontal.

The geostrophic estimate of the PV, as shown in Eq. (11), involves the difference between the vorticity and the inverse Richardson number, and both these parameters are large in the frontal zone; therefore the magnitude of their difference is difficult to estimate simply from either of the component fields. Also this difference is multiplied by the static stability which also shows an anomaly in the frontal zone. It is clear from this discussion that it is not straightforward to deduce the potential vorticity distribution on the basis of either the vorticity or the static stability fields. From the discussion in Hoskins et al. (1985) it can be shown, however, that a narrow but deep anomaly of PV will be mostly composed of a vorticity anomaly, whilst a broad and shallow PV anomaly will be mostly associated with static stability changes.

In Fig. 13 several examples of the PV distribution at fronts are given. A striking feature is the large amplitude, but mesoscale, positive anomaly in the lower frontal zone. This is in accord with the theoretical ideas discussed earlier. The maximum value is about 1.5 PVU, which is otherwise typically found at the tropopause level. The maximum is at about 850 mb in the frontal zone so is unambiguously not due to the descent of tropopause PV but rather due to the latent heating occurring in the saturated ascent in that region.
Figure 12. Cross-sections showing the distributions of $N^2$ (contour interval $1 \times 10^{-4}\,\text{s}^{-2}$ and values greater than $1 \times 10^{-4}\,\text{s}^{-2}$, shaded), absolute vorticity (contour interval $1 \times 10^{-4}\,\text{s}^{-1}$ and positive values shaded) and Richardson number (contour interval 0.5 for values less than 2, and values less than unity shaded) given from run 3 of IOP 7. The sloping frontal zone shows clearly as a region of enhanced stability and vorticity and with a low Richardson number.
Figure 13. The potential vorticity distributions for run 3 of IOP 3, 7, and 8. The contour interval is 0.5 PVU with positive values shaded. The near-surface maximum in excess of 2 PVU is evident in each case, as is the stratospheric intrusion at upper levels (in IOP 3 there is a mid-tropospheric maximum which is not of stratospheric origin). Note the regions of negative values in the warm air.
of the front. This raises the difficulty of interpreting tropopause exchange solely in terms of the \( PV \) distribution. The troposphere, in the vicinity of fronts, is a region of inherently anomalous \( PV \) owing to diabatic and frictional processes. The maximum value of \( PV \) is about 5 times that of climatology at 850 mb (see Hoskins et al. 1985), whilst the frontal vorticity may be 3 times \( f_0 \); these figures give an impression of the relative contributions to the \( PV \) from the vorticity and the static stability.

As discussed earlier, patterns of potential vorticity provide an alternative means to cross-sections of \( \theta \) (or \( \theta_e \)) and momentum for study of symmetric instabilities. Although rarely used for mesoscale studies because of limited data resolution, we quote them here because they make evident coherent features common to several runs and several cases more clearly than the undifferentiated variables. Consideration of these patterns also leads to pointers to the processes causing instabilities.

Regions of negative potential vorticity are frequently evident in certain locations, particularly in the warm sector at mid-tropospheric levels; this was particularly marked and well resolved in the IOPs 7 and 8. From Fig. 12 it can be seen that the absolute vorticity is also negative in much of this area. This important feature of frontal structure has been previously commented on by Roach and Hardman (1975). As the regions are also convectively stable, even dry symmetric instability appears possible throughout a substantial region. Interestingly, the most markedly negative values occur in run 4 of IOP 7 in the region of the pronounced cross-frontal circulation noted earlier. It is known from simulations of CSI that the action of the unstable roll-circulations is to generate local regions of convective and inertial instability. It may be that the observed regions of negative absolute vorticity are a response to CSI and the release of the implied inertial instability is inhibited by the CSI and frontal circulations. However, this feature is one which clearly needs further theoretical research.

Figure 14 shows cross-sections of equivalent potential vorticity corresponding to those for \( PV \) in Fig. 13. Values near zero clearly dominate the field, in accordance with current theories and observations, but again negative values are widely distributed about the front. Broadly these arise in the conditionally unstable layer over the warm sea surface, but also they occur in the statically stable warm sector aloft. However, the most remarkable feature is the pattern of sloping patches, in places intensely unstable, lying across the frontal zone between these two regions. Such patterns were completely unanticipated in any previous theoretical or observational studies. It should be noted, however, that often in the lower troposphere intense precipitation is associated with positive anomalies of \( PV_e \) which is not consistent with the CSI mechanism operating in these locations.

Our earlier theoretical discussion suggested several possible sources of equivalent potential vorticity. The low-level region of negative values can clearly be explained by fluxes of moist entropy from the sea surface. The only likely source of negative values aloft appears to be the occurrence of three-dimensional unsaturated motion. Clearly study of this aspect is outside the scope of the present work, but three-dimensional objective analyses will be given in subsequent studies of individual cases. The presence of negative values over an extensive region in the warm sector at mid-tropospheric levels, as in IOP 7, is difficult to explain in a nearly two-dimensional environment (using Eq. (4) to estimate the production rate). Thus advection of air from a mature storm system is the likely source for this feature rather than \textit{in situ} generation. The existence of the sloping patterns below, however, suggests a locally intense sink, which we suggest is evaporation at the rear edge of the cloud mass. This process is indeed three-dimensional since the intensity of precipitation is strongly modulated over a period of hours by the passage of frontal waves.
Figure 14. As in Fig. 13 but showing the equivalent potential-vorticity distributions. Note the extensive regions of negative values along the front and in the warm air. Also note the tendency for variations lying across the frontal slope.
The high resolution of the dropsonde observations clearly challenges the conventional view that there is even small stability to symmetric motion at fronts.

\[(j)\] **Three-dimensional structure**

Although the aim of the experiment, and choice of cases, was to observe two-dimensional systems as far as possible, nearly all the fronts exhibited appreciable frontal-wave activity. The sounding strategy was not best suited to study the waves themselves, which appear to have a horizontal wavelength in the along-front direction of perhaps 700–1000 km. They were relatively weak features in terms of surface pressure, with anomalies of only a few millibars. However, they are evident in variations of the frontal zone on satellite pictures and in differences between the 4 runs of each event. Variations between the runs of a given front are consistent with waves being local enhancements and diminutions of the vertical velocity, with the attendant changes to the horizontal flow. Given the emphasis in this paper on the importance of diabatic processes in such fronts, the frontal wave, being a region of enhanced ascent, cloud, and precipitation, is an important feature in modifying the dynamics. For example the local changes to the flow near the sea surface will produce a change in surface fluxes and consequently modify convection also. Further studies will consider the operation of such mechanisms in individual systems, with the aid of numerical models.

5. **Discussion**

In this paper we have identified certain common features of active cold fronts which are evident in the high-resolution dropsonde data. Structures with horizontal scale of between a few kilometres and a few hundred kilometres have been described from these data; these cannot be defined either by routine radiosonde or by regional forecast model output data. The discussion has not followed the traditional lines of presenting a 'complete' picture of individual events; rather the focus has been on analysis of the dynamical variables postulated in the pre-experiment hypotheses as being critical for the verification of theoretical models of fronts. Now we re-consider those hypotheses that relate to the mesoscale structure and draw together the main conclusions from this paper in each case. Of the eight hypotheses quoted in section 2, numbers 2, 3, 5, 6, 7, and 8 have been addressed herein, with numbers 2, 3, and 5 being of particular interest.

**Hypothesis 2.** *Frontal motion is balanced...*. This notion, which is really central to the semi-geostrophic model of frontogenesis, has been assessed by analysing the thermal wind imbalance. Zones along the sloping front of substantial thermal wind imbalance have been observed, but they are seldom of horizontal extent greater than about 50 km. There is therefore no evidence to support the idea that frontogenesis may be limited by the onset of inertia-gravity-wave activity associated with ageostrophic accelerations. Such waves were observed during the frontal passage by using sensitive microbarographs but they appear not to disrupt frontogenesis.

**Hypothesis 3.** *Neutrality to slantwise moist ascent...*. Perhaps for the first time these observations have provided direct evidence of conditional symmetric instability occurring at fronts. This is revealed by the existence of buckled \( m \)-surfaces in the lower troposphere. In such zones, which are limited to horizontal scales between 50 and 100 km, neutral conditions certainly do not apply. However in the mid-frontal and upper-frontal zone, even in regions without saturation, there is a remarkable parallelism of \( m \) and \( \theta_{es} \) surfaces probably indicating an adjustment to near-neutral conditions to CSI. Also observed on
occasion is a tendency for the 'absolute momentum' to be well-mixed in the warm air just ahead of the front in the lower troposphere. This may be linked to slantwise adjustment of air exiting from the top of line convection at the front.

**HYPOTHESIS 5.** 'Diabatic processes are a substantial influence...' There is strong evidence in these data that diabatic processes are controlling many aspects of the frontal dynamics. We cite two examples now: the first is the large anomalies of potential vorticity in the front, almost certainly due to diabatic processes, and the second is the substantial enhancement of the descent under the frontal surface caused by snow evaporation.

**HYPOTHESIS 6.** 'Surface cold front behaves like a density current'. Whilst evaporative cooling is observed in the cold air, the mesoscale and synoptic dynamical constraints would appear to rule out describing the propagation of the cold front with density-current dynamics on the scale of these observations.

**HYPOTHESIS 7.** 'Frontal waves associated with PV anomalies'. All the fronts observed in this experiment were, to quote the forecaster's word, "complicated" by the presence of relatively small amplitude frontal waves moving rapidly toward the low centre. From these data it is not possible, given their typical scale of several hundred kilometres, to state whether they owe their existence to either upper or lower tropospheric potential-vorticity anomalies. That such anomalies are ubiquitous can be taken as direct evidence that this hypothesis may not be refuted from these data.

**HYPOTHESIS 8.** 'Most convergence in the boundary layer'. In nearly all cases there was a substantial convergence in the lowest kilometre into the frontal zone from both the warm and the cold air. The structure and role of the highly disturbed boundary layer in frontal zones requires further attention with small-scale observations and theoretical studies. However by no means all frontal convergence can be accounted for in this way. Alternate layers of convergence and divergence (of comparable magnitude) sloping along $\theta_c$-surfaces, were a notable feature of the observations throughout the troposphere.

An important extra feature, which was also observed at a warm front in Project Scillonia, is the existence of substantial regions of negative absolute vorticity. These occur not only on the anticyclonic side of the upper jet but also in the lower troposphere. This latter region appears to be symptomatic of the inefficiency of the dynamical processes in releasing inertial instability, compared to the rapid generation of negative absolute vorticity by moist processes such as CSI.

It is clear that we have laid great emphasis on the role of the variables which are conserved in various regions of the front. These are:

- normalized absolute angular momentum (or 'absolute momentum');
- equivalent potential temperature;
- potential vorticity;
- equivalent potential vorticity.

The use of potential vorticity in describing the mesoscale structures at fronts is central to the understanding of their dynamics. The extent to which this is true is governed by the extent to which along-front geostrophy, the main plank in the semi-geostrophic theory, is well justified. But not included in that model, until the recent work of Emanuel et al. (1987) and others, are the substantial changes to the potential vorticity by the diabatic processes at fronts, as observed in the data presented here. This does not invalidate the semi-geostrophic model but does require the different viewpoint presented here. The use of potential vorticity is, in principle, also capable of unifying the semi-
geostrophic model and the so-called conceptual or, more accurately, the airflow model of fronts. This synthesis has not been attempted here but it is the next important task in frontal-dynamics research.

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