Dynamical effects of ice sublimation in a frontal wave

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SUMMARY

The dynamical role of ice sublimation in weather systems is briefly reviewed. Observations are presented from the Fronts and Atlantic Storm-Track Experiment (FASTEX) Intensive Observation Period 16 that show variations of static stability, humidity and mesoscale circulation corresponding to those associated theoretically with sublimation of ice precipitation. It is thus suggested that the observations display the mechanism proposed by Clough and Franks in which forward mesoscale flows are associated with moist adiabatic descent supported by the sublimation cooling. This mechanism was suggested as an important stage in the evolution of many mesoscale rain bands.

A set of three model simulations of the event has been made with versions of The Met. Office’s Unified Model. Of these a mesoscale model integration with 11 km resolution and 45 levels clearly displays the symptoms, and is diagnosed to demonstrate its consistency with the Clough–Franks mechanism. An integration omitting the cooling due to sublimation differs significantly from the full model experiment in the structure of low-level wind fields, frontal troughs and mesoscale precipitation distribution. It is also demonstrated that the static-stability transition, mesoscale circulation and mid-tropospheric potential-vorticity perturbations are substantially weakened in this integration, thus confirming that the Clough–Franks mechanism is also operating in the numerical weather prediction (NWP) model.

We deduce from these studies that ice precipitation and its sublimation has a major role in determining mesoscale circulation and structure in mid-latitude weather systems, affecting stratification and the formation of features such as fronts and rain bands. These are substantially affected by the fall and evaporation of ice crystals, which are both important to temperature and moisture transports and the behaviour of NWP models on timescales of hours to days. In our integrations dynamical feedback due to sublimation cooling coincided with extreme negative potential-vorticity values and potentially conditional symmetric instability, hence the anticyclonic motion occurring in the cloud head or deep cloud of the moist warm sector as in this case. In moist warm sectors a substantial role for sublimation may be anticipated more generally, particularly for air trajectories receiving most ice precipitation.

We suggest that the described phenomenon be referred to as sublimation enhanced descent or SED.

It is concluded that in view of this demonstrated sensitivity substantial attention should be given to refining microphysical parametrizations in NWP models, and that radar and sounding observations from the FASTEX experiment provide a suitable basis for validating these schemes.

KEYWORDS: Atmospheric dynamics Microphysical processes Numerical weather prediction

1. INTRODUCTION

Studies of cloud processes, forecasting and weather system dynamics have long been regarded as distinct specializations within meteorology. Such distinctions do not exist in the atmosphere, and the ascendancy of numerical modelling is forcing us to transcend them. Two particular examples are important in the present context.

Firstly, most of the structural features evident in satellite imagery from active weather systems are produced by the formation and dissipation of ice clouds. Yet these are commonly interpreted in terms of forecasting or dynamics, with no reference to the properties of ice crystals and the microphysics of their formation or fate. Secondly, with the exception of Clough and Franks (1991; henceforward CF) current theories of the development of mesoscale rain bands developed from dynamical ideas that required only latent-heat release regardless of details of cloud and precipitation microphysical processes.

Modern numerical weather prediction (NWP) models, however, contain complex parametrizations describing putative cloud microphysical and dynamical properties.

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It is essential for future progress that these models be interpreted within a physical framework which embraces the basic physics of the atmospheric processes represented. This is the central purpose of the present work.

Suppose we consider the cloud and precipitation processes as a hierarchy ordered by their importance in the latent-heating budget of a system. Then the leading term, release of latent heat on ascent, is indeed positive and independent of microphysics (to the extent that the modifying effect of glaciation is ignored). This is part of the reason why NWP models have for so long been capable of forecasting large-scale precipitation with little attention to microphysics. The other part is that the microphysical processes of ice are actually so fast that they permit highly efficient ice crystal growth by diffusion; the dynamics of water-vapour supply is thus the rate-limiting step (Wexler and Atlas 1958). This is evident in that most of the time weather system precipitation is crystalline rather than heavily rimed, because the rate of crystal growth is rapid enough to keep pace with the supply of supersaturation at rates of mesoscale ascent of up to around 30 cm s\(^{-1}\).

The next most important term in mid latitudes is the cooling of the atmosphere by sublimation, i.e. evaporation of ice precipitation, which can in principle locally exceed the potential for latent-heat release. Unlike the latent-heat-release term, however, the sublimation rate is highly sensitive both to particle characteristics and to ambient conditions. Sublimation is also very fast, since diffusion growth of crystals is almost reversible at microscopic scales; commonly particle mass 1/e-folding times for sublimation are of order 10 minutes, corresponding to depth-scales of hundreds of metres. However, because ice crystals interact so strongly with their environment, the availability of precipitation and variation of thermodynamic state in the fall path become critical.

This complex and interesting behaviour of sublimation has received very little attention in the literature. Two paradigms have been suggested. In the first, Harris (1977) attempted to explain downdraughts of several metres per second in Doppler radar observations by sublimation into the sub-cloud layer. He successfully used microphysical model calculations to show that the layer immediately beneath cirrus cloud can be cooled so intensely that static instability may eventually result and hence lead to dry convection. A second situation was envisaged by CF, who suggested that the detailed physics of sublimation is likely to play an important role in the occurrence and structure of broad (10–100s km) rain bands or snow bands. From simple kinetic–thermodynamic models they showed that in stable conditions sublimation should be capable of maintaining moist slantwise descending mesoscale circulations of up to 20–30 cm s\(^{-1}\) in ice precipitation, typical of those occurring in many rain bands. Such moist slantwise descent, they argued, is an essential feature of spatially extensive multiple bands produced via conditional symmetric instability (CSI; Bennetts and Hoskins 1979) or a similar mechanism.

The above two scenarios are quite different, though in some respects complementary, and may co-exist within a weather system. The latter, however, has a clear implication for mesoscale weather system dynamics. The prediction of slantwise descent has been confirmed qualitatively by Marecal and Lemaitre (1995) in Doppler radar observations and model calculations showing the importance of subliming snow in the mesoscale descent zone of an observed rain band.

The Fronts and Atlantic Storm-Track EXperiment (FASTEX) presents the ideal opportunity to pursue this problem quantitatively by providing collocated dropsonde, Doppler radar and aircraft microphysical observations capable of validating mesoscale model predictions (Joly et al. 1997). During January and February 1997 this combination of observing platforms was used to characterize the mesoscale structure
of several mid-latitude frontal systems in unprecedented detail. In a number of Intensive Observing Periods (IOPs) aircraft equipped with Doppler radar and dropsondes flew together along adjacent parallel tracks, so that both clear and precipitating cloudy regions were observed over distances of 400–800 km, removing many of the limitations of coverage normally encountered in the study of weather systems.

Of the events observed, IOP16 is best suited to the present study because of the simplicity of both the weather system and flight pattern. A rapidly moving trough intensified to form a low-pressure centre during the period of the aircraft dropsonde and radar observations. The aircraft flew six equally spaced tracks nearly perpendicular to the track of the weather system. This pattern permits model validation in unparalleled detail, and ultimately should permit comparison with microphysical process rates systematically calculated from the complementary observational datasets.

In this study we draw attention to a number of observed features of IOP16, which we take to be symptomatic of sublimation and the associated slantwise descent. As well as thermodynamic structure, these include dynamical effects indicating mesoscale circulations driven by the cooling, according to the mechanism of CF. A number of mesoscale model simulations of the event are then presented in which comparable features are indicated and the role of sublimation demonstrated. It is argued that this role is primarily to reduce the effective static stability to modest descent towards the moist value, and hence affect the response to local dynamical forcing. This mechanism implies a clear active role for ice precipitation in the mesoscale dynamics of weather systems, which must be appreciated in order to develop our quantitative understanding beyond the limitations of current conceptual models.

In section 2 we briefly review the sublimation process and the key results of the earlier works of CF and Harris (1977), with emphasis on the relevant time- and depth-scales. Section 3 provides an overview of the development of the system observed in IOP16 and its essential dynamics. In section 4 we show the mesoscale symptoms of sublimation in analyses of dropsonde observations, while section 5 presents the results of the model simulations. In section 6 we propose a hypothesis extending our results more generally to the effects of sublimation on frontal waves. In section 7 we summarize our results and discuss their implications both for NWP and for our understanding of weather system structure.

2. Sublimation Effects in Weather Systems

(a) Process rates

The microscale physics of sublimation is simple and well established. However its operation in the atmosphere is dynamically and thermodynamically interactive and poorly documented. A number of key features can be deduced from thermodynamic and kinetic principles. In this section we discuss these principles to give some appreciation of the time- and depth-scales of sublimation without recourse to detailed calculation. This will prove a necessary background to our later observational discussion.

Microscopically, sublimation is the transfer of molecules from the ice crystal lattice to the gas phase, the reverse of the condensation process by which crystals grow directly from the gas phase (Pruppacher and Klett 1978). Both processes are given by the following equation for the rate of change of mass of an ice crystal with time:

\[
\frac{dM}{dt} = CF \frac{S}{(L/KT)(L/RT - 1) + RT/eX}.
\]
The rate equation is represented as a product of three terms: the capacitance, $C$, the ventilation factor, $F$, and the environmental factor. $C$ is solely a function of ice crystal geometry. $F$ is typically of order 1–3 and represents the enhancement of the diffusion mass growth rate when ice crystals are falling in the atmosphere; it depends on crystal geometry, fall velocity and the ambient air density. The third term represents the effect of the environment on the growth rate and is independent of crystal geometry. The diffusion of water vapour towards the ice crystal and the thermal conduction of latent heat of sublimation away from the crystal are both represented in this term, where $S$ is the supersaturation of water vapour in the environmental air with respect to ice, $L$ is the latent heat of condensation or sublimation, $T$ is the temperature of the environmental air, $e$ is the saturation vapour pressure with respect to ice, $X$ is the diffusivity of water vapour in air, $K$ is the thermal conductivity of air, and $R$ is the universal gas constant.

While $T$ and $e$ are important factors in determining the ice crystal growth rate, atmospheric variations tend to be dominated by $S$.

Despite the low temperatures diffusion is rapid, because it involves molecular transfer over microscopic distances with only a minor energy barrier due to intermolecular forces. Thus sub-zero precipitation in frontal weather systems is normally crystalline, with a variable degree of riming because the rate of condensation growth of ice is roughly sufficient to balance the provision of supersaturation by mesoscale ascent, which is typically of order 20–30 cm s$^{-1}$ or less (Rutledge and Hobbs 1983).

Potentially the sublimation of snow can proceed more rapidly than its growth by diffusion. This can be seen by considering the limitations imposed by humidity. The upper limit on the subsaturation is 100%, whereas the supersaturation with respect to ice in the atmosphere is limited by water saturation, a smaller amount varying between 0% at 0 °C and 47% at $-40$ °C, as shown in Fig. 1.

Further, as CF remark, since the saturation vapour pressure increases with temperature, the potential cooling due to sublimation must also reach a maximum near 0 °C. For modelling this has the significant implication that representation of the correct phase of cloud and precipitation in the range 0 to $-10$ °C is important for accurate calculation of the effects of sublimation in the atmosphere.

The above arguments indicate the potential rapidity of sublimation. Precise particle calculations indicate that the 1/e-folding time for typical ice crystals is of the order of minutes at 60% saturation (CF). It is much shorter than the time-scale for evaporation of raindrops because of the increased surface area of ice particles of the same mass, typically by a factor of 3–5. CF studied this very different atmospheric behaviour in detail and showed that much of it resulted from the different particle and distribution characteristics of snow and rain. The evaporation of rain at modest rainfall rates is inhibited by the loss of the small drops which contribute most to the mass loss. This does not occur for snow because the bulk density of snow is approximately proportional to $1/D$ for particles of diameter $D$ (Locatelli and Hobbs 1974), hence the mass loss is uniformly distributed across the spectrum, of which small near-stationary ice crystals represent a small component.

The dynamical importance of precipitation evaporation is given by the latent-heat loss, which is approximately proportional to the mass that evaporates in a given depth. CF express this by the depth-scale for 95% evaporative mass loss, a shallow depth-scale implying the greatest effect. This scale is inversely proportional to the rate of mass loss above multiplied by the residence time ($\sim 1/\text{terminal velocity}$). The long residence time of snow compared to rain (for typical mean terminal velocities of 1 m s$^{-1}$ compared to 3–5 m s$^{-1}$) further emphasizes its importance.
The bulk-density and terminal-velocity effects thus combine to give depth-scales for snow sublimation of hundreds of metres compared to evaporation distances of several kilometres for rain or graupel, and so sublimation is much more important than rain or hail evaporation in frontal-wave dynamics (and in middle latitudes generally except in very heavy rain). This shallow depth-scale is probably also an important factor in the formation of multiple layers of cloud and associated temperature and moisture fluctuations frequently observed in frontal systems (GEWEX 1996)

(b) Thermodynamic and dynamical effects of sublimation

When saturated air parcels rise, latent heat of condensation is released immediately into the air in moist adiabatic ascent. If an air parcel laden with precipitation were to descend, then that process would also be moist adiabatic if the descent rate were small enough for evaporation or sublimation to keep pace. The process would occur spontaneously, being thermodynamically reversible. The critical question, then, is one of kinetics: how rapidly in practice can an air parcel descend while remaining essentially saturated?

As noted in the introduction, CF indicated that for sublimation this descent rate is potentially 20–30 cm s$^{-1}$, enough to support typical mesoscale circulations. The critical point to note, however, is that the sublimation cooling feedback is spontaneous, and mesoscale descent will tend to follow the moist adiabat while there remains ice precipitation to evaporate.

The process of sublimation, then, reflects a tendency to maintain saturation with respect to ice despite the drying influence of subsidence warming in stable environments; in neutral to unstable conditions mixing and overall drying in terms of relative
humidity results, as in Harris (1977). Also, the Bergeron process presents a tendency of the atmosphere to attain ice saturation rather than water saturation in ascent above the melting layer in weather systems. Figure 1 shows the distribution of relative humidity with respect to ice versus temperature for all of the Vaisala GPS dropsondes used by The Met. Office's C-130 aircraft in FASTEX. The concentration of observations near 100% clearly indicates the powerful influence of the ice deposition/sublimation mechanism. It should be noted that corresponding traces (not shown) using data from earlier FRONT5 87 (Thorpe and Clough 1991) and FRONT5 92 (Browning et al. 1995) dropsondes with carbon hygristor moisture sensors do not present such a pattern, which we attribute to their lower accuracy and reproducibility.

The next issue to consider is the environmental constraints operating on a sublimation-cooled, and therefore denser, air parcel. Clearly the static stability is likely to be an inhibiting factor in general. Harris (1977) considered the local situation of ice streams evaporating into a near-neutral environment. Without dynamical feedback the cooling is clearly greatest immediately beneath the precipitating cloud, and so a stable layer results above a layer with reduced stability. As precipitation continues the lower layer is progressively destabilized until it becomes super-adiabatic and convection driven by cooling breaks out. Harris demonstrated that this could occur through a 400 m layer within 1 h with typical precipitation rates of the order of 0.1 mm h⁻¹. Downdraughts of several m s⁻¹ were predicted by a one-dimensional dynamical model, which showed significant dependence upon the assumed particle characteristics.

In these neutral–unstable conditions air parcels, paradoxically, are dried by the sublimation, in the sense that their relative humidity is reduced by the ensuing descent. For such air parcels within frontal environments this process might progressively produce deep layers of near-neutral air. The determining factor is the time spent by such parcels in regions of substantial ice precipitation rates.

In stable frontal environments descent can most readily occur in slantwise trajectories, in frontal zones where latent-heat release has reduced the moist potential vorticity and hence absolute momentum, and wet-bulb potential-temperature, θ_w, surfaces are near parallel (Bennetts and Hoskins 1979). The most evident observational features characterizing this behaviour are strong localized forward cross-frontal circulations beneath frontal cloud in a sense corresponding to direct cooling (CF; Thorpe and Clough 1991). The slantwise-descent mechanism envisaged by CF is shown schematically in Fig. 2. From this it is evident that the relative humidity with respect to ice is a key indicator in principle, since it must be slightly subsaturated in order to cool and produce descent. (Note also that the relative humidity is thus seen to be a consequence of the descent and ambient precipitation rate, rather than itself a determining quantity.) Unfortunately, in practice the accuracy even of these new humidity measurements is insufficient to meet the stringent 1–2% absolute accuracy necessary to test the small subsaturations of order 5% likely where the CF mechanism operates. A better check is provided by gradients of radar reflectivity indicating the depletion of precipitation, since they must occur through the same zone as the forward descent.

The role of terminal velocity is vital in determining the depth-scale and intensity of sublimation effects. The primary effect of latent heating in weather systems is to produce a low-level maximum of potential vorticity, which acts as a source for increased vorticity near the surface (Hoskins et al. 1985). By imposing a layer of cooling beneath the layer warmed by latent heating, the effect of sublimation is to create a pattern which may be expected to resemble a shallow but intense vertically oriented dipolar distribution of heating. The corresponding modification of the distribution of potential vorticity, from attribution principles, may be expected to have rather localized response, primarily in
static stability rather than vorticity (Hoskins et al. 1985), and so little direct effect might be expected in the surface pressure pattern (Parker and Thorpe 1995).

Clearly the above precipitation effects are complex and involve substantial dynamical-thermodynamic interaction at the mesoscale. In view of this complexity there is a clear deficiency in the modelling literature in this area; much more information may well be needed to adequately parametrize the range of interactions and possible scenarios.

3. The Event

The frontal wave studied in IOP16 formed from a weak trough on the main baroclinic zone in the western Atlantic. It passed Newfoundland around 0600–1200 UTC 16 February 1997 and travelled rapidly (30–40 m s\(^{-1}\)) across the Atlantic, deepening to form a low pressure centre in the first half of 17 February.

Figure 3 shows the development characterized in three-hourly Meteosat infrared geostationary satellite images for the observation period, overlaid with contours from The Met. Office’s operational 50 km Limited-Area Model (LAM) analyses. The Met. Office’s operational LAM forecast beginning at 1800 UTC 16 February was used for planning the flights to obtain aircraft-based observations. This forecast was consistent with its predecessors and proved accurate.

The trough deepens from 0300 UTC to form an emerging cloud head, which is clearly distinct from the upper-cloud shield by 0600 UTC as shown in Fig. 3(b). Also in the interval the uniform rear part of the cloud shield thins in places to result in a pronounced set of bands, some of them parallel to the rear edge by 0600 UTC, clearly evident in Fig. 4 which is an enlarged section of Fig. 3(c) west of Ireland. From this stage the wave develops rapidly, forming a secondary cloud head between 1000 and 1100 UTC (Fig. 3(d)). The LAM indicates the formation of a closed low centre near 1200 UTC, though the soundings suggest that it probably occurred around 0900–1000 UTC in the atmosphere.
Figure 3. Sequence of 3-hourly Meteosat infrared images for 0300 UTC to 1200 UTC 17 February 1997, FASTEX IOP16 (see text). The black contours are surface pressure from The Met. Office’s Limited Area Model forecast starting at 1800 UTC 16 February 1997, with contours at intervals of 8 hPa. Grey shades range from lightest $< -54 \, ^\circ \text{C}$ to darkest $> +18 \, ^\circ \text{C}$ in 6 degC intervals. The cloud head starts to emerge around 0300 UTC. The dashed line on the 0900 UTC image shows the flight track of the UK C-130 aircraft spaced out in a frame relative to the velocity of the moving low pressure system at around 0900 UTC.
Figure 4. Meteosat infrared image for 0600 UTC 17 February 1997, an enlarged section of that shown in Fig. 3(c). The locations of drop sondes for the first run (see text) are indicated by drop sequence numbers.

The C-130 and P3 aircraft worked essentially to their pre-flight plan to map out six runs perpendicular to the wave’s path and separated by about 100 km along the wave, assuming a system speed of $28 \text{ m s}^{-1}$ from $240^\circ$. These are indicated in Fig. 3(c), when displaced according to the system velocity; 62 drop sondes were deployed, of which 45 gave usable data. Important features from the first run are discussed here; sonde locations are shown on the 0600 UTC satellite image in Fig. 4.

4. OBSERVED CHARACTERISTICS

(a) Data

The soundings were made using 45 RD82 GPS drop sondes developed by Vaisala and flight-tested by The Met. Office. The temperature and humidity elements were capacitive sensors as in the standard RS80 radiosonde, while the pressure sensor was the recently developed Barocap silicon pressure transducer as in the new RS90 radiosonde. Winds were provided by the Vaisala GPS111 receiver module on the sondes with Vaisala NWG201 processor cards on the C-130.

The humidity observations appear much more consistent than those from earlier FRONTS 87 and FRONTS 92 dropsonde observations. This judgement is based upon the increased density of relative-humidity observations near the ice saturation curve shown in Fig. 1, which is qualitatively a reasonable property to expect for such environments where the Bergeron–Findeisen process is operating. Despite the improved
consistency, however, the high frequency of values above ice saturation suggests that the observations may be about 5% too high in the moist part of the range.

The basic parameters from the dropsonde data (temperature, scalar wind components and humidity) were objectively analysed as two-dimensional vertical cross-sections for each of the runs indicated in Fig. 3(c) (Forbes et al. 2000a). The objective analysis uses a successive-correction algorithm (Pedder 1993) with isentropic versus horizontal distance coordinates. Data are extracted on isentropic levels every 0.5 degC from each sounding for input to the objective analysis. Isentropic coordinates are natural for frontal zones as the correlation along such surfaces is generally higher than along pressure surfaces, i.e. frontal flows tend to follow isentropic surfaces.

It should be noted that filtering is inevitably necessary in objective analysis to remove scales which cannot be represented adequately by the distribution of observations. Thus the horizontal-filtering-scale used in the objective analysis is approximately twice the mean sounding separation (about 60 km). The vertical-filtering-scale must also be defined appropriately so that only features with resolved horizontal-scales are resolved in the vertical, otherwise fragmented structures arise in the analyses. Some loss of information in the original soundings is therefore inevitable. The isentropic-filtering-scale used in the analysis is 1.5 degC, corresponding on average to a vertical-scale of around 30 hPa. With this scale some features, for example shallow potentially unstable layers in the original sounding data, are lost, but the resulting cross-sections are more nearly comparable with those from numerical models. Horizontal distances in the figures are shown as the distances projected onto the plane chosen to resolve vector quantities, perpendicular to the system velocity.

(b) Run 1

The location of run 1 is shown in Fig. 4. The authors’ attention was originally drawn to the characteristics of this run by the fact that the main observed frontal features discussed below were totally absent from the operational 19-level LAM forecast, despite the fact that the model correctly forecast the overall development and pressure fall. Since a good forecast is available the case is particularly valuable for investigating the sensitivity to detailed processes.

Harris’s (1977) study centred on the formation of a cooled layer at the base of an upper-level cloud. A frontal surface is of course a typical location of such a static-stability transition, and a well-defined example occurs in run 1, though unlike in Harris’s case the cooling is not sufficient here to generate the convection which was the focus of his study. This may be deduced from Fig. 5, which shows cross-sections of key thermodynamic and kinematic quantities and radar reflectivity.

The cross-section of \( \theta_w \) (Fig. 5(a)) shows a modest low-level gradient, but a particularly strong and coherent gradient is evident across several soundings between 450 and 550 hPa, corresponding to a frontal surface. This may be seen more fully in the dry static stability, the square of the Brunt–Väisälä frequency, \( N^2 \), in Fig. 5(b).

Beneath the stable frontal layer in Fig. 5(b) is a substantial layer of reduced static stability around 550 hPa, becoming moist unstable in a shallow layer beneath sondes 6 and 7 in the original soundings (the shaded region in Fig. 5(a)). At full data resolution soundings 8, 9 and 11 also possess shallow dry-adiabatic layers immediately beneath the inversion, a typical symptom of intense sublimation cooling. The extent of the upper-stability transition is striking, crossing 300 km of the section from sonde 1 to sonde 8 along the 13 °C \( \theta_w \) surface. It is suggested that sublimation from the cloud layer above makes an important contribution to the presence and amplitude of such a structure.
The section-parallel flow shows particularly interesting structure (Fig. 5(c)). A shallow layer of strongly descending forward flow in mid troposphere coincides with much of the stable zone of the upper front. The maximum flow from 450 to 550 hPa at section-parallel distance $x = 100$ to 450 km, and a lesser maximum at 650 to 700 hPa, $x = 300$ to 450 km, occur beneath regions supersaturated with respect to ice in Fig. 5(d) and coincide with upward gradients of relative humidity. The reflectivity pattern in Fig. 5(f) confirms the presence of moist precipitating regions. The supersaturated regions correspond approximately to reflectivity values of up to $-10$ to $-5$ dBz, typically precipitation rates of about 0.025 to 0.05 mm h$^{-1}$, which decrease steadily through and beneath the forward descending flow.

This feature is well above the melting level (Fig. 5(a)), where ice precipitation processes clearly dominate and the situation envisaged by CF must apply, and the pattern is qualitatively consistent with that predicted by the Clough–Franks mechanism as in Fig. 2. Unfortunately, because of the above stringent requirements and the evolving characteristic of such situations, quantitative testing is not practicable with atmospheric humidity data. We thus find the confirmation of symptoms in a dynamical model, as in the following sections, to be a more practicable confirmation that the mechanism is operating.

The overall forward flow in the warm sector aloft is associated with the anticyclonic flow into the ridge, strengthening downstream of the main latent-heating zone. This corresponds to the reduction of potential vorticity which is expected above and downstream of the level of maximum latent heating; this is also consistent with model diagnostics to be shown later. The local maximum in the frontal surface corresponds to a direct circulation driven by local sublimation cooling, and the fact that the maximum coincides with the lower part of the saturated region is consistent with CF’s proposal of moist adiabatic descent occurring where precipitation is sufficient to support it. The layer is shallow because the available precipitation sublimes efficiently but is insufficient to support a deep layer of descent, hence the structure degenerates into finer scales because of the moist processes.

Figure 5(e) shows the momentum, $v + f x$, where $v$ is the section-perpendicular velocity component and $f$ the Coriolis parameter. From theory the shallow slope of the momentum surfaces through the upper front is commonly associated with a low potential-vorticity (or moist potential-vorticity) value where momentum and potential-temperature, $\theta$, or $\theta_{d}$ surfaces tend to become parallel and symmetric instability (SI) or CSI may occur. The two-dimensional estimate of potential vorticity is useful, though not absolutely reliable in the curved flow, but we note that the main region of shallow slope coincides with the frontal regions of both high and low static stability. In this case the shallowest slopes coincide with strong moist static stability (cf. Fig. 5(a)) and the strong section-parallel flow in Fig. 5(c). Beneath this, between 230 and 340 km at 550 to 600 hPa is a limited dry region of negative moist potential vorticity in the two-dimensional estimate (not shown). Whilst noting the occurrence of this region, the strongest circulation coincided with the more stable region above, suggesting the sublimation source to be a more important factor for the mesoscale circulation here than CSI. Indeed, since the location of the low moist potential vorticity is closely below the region we suggest to be strongly cooled, sublimation may in this case give rise to the implied instability; this possibility is considered in the later discussion of our model results.

The deepest area of weak static stability between sondes 6 and 8 corresponds to a gap between cloud bands in Fig. 4, which is also evident as an absence of ice particles in the aircraft data, with concentration maxima on either side. The strong horizontal gradient
Figure 5. Cross-sections of observed quantities from run 1 (see text). Distances (km) are indicated projected onto an axis at 140° through the sondes. Dropsonde locations are indicated on the top axis. (a) Solid contours of wet-bulb potential temperature $\theta_w$ (°C); the shaded region has become moist unstable (see text), while the dashed curve is the 0 °C isotherm. (b) The square of the Brunt–Väisälä frequency (s$^{-2}$); heavy shading indicates values greater than $2 \times 10^{-6}$; light shading indicates values less than $1 \times 10^{-6}$. (c) System-relative section-parallel wind (m s$^{-1}$); light shading indicates values greater than 4 m s$^{-1}$; heavy shading indicates values greater than 8 m s$^{-1}$. (d) Relative humidity (%) with respect to ice: light shading indicates values less than 96%; heavy shading indicates values greater than 104%. (e) Momentum, $v + f x$ (m s$^{-1}$; see text). (f) Reflectivity (dBZ) from National Oceanic and Atmospheric Administration P3 aircraft Doppler radar observations.
in the section-parallel velocity (Fig. 5(c)) in the sloping region AB from $x = 150$ km, $p = 700-500$ hPa to $x = 350$ km, $p = 800-700$ hPa, is consistent with the occurrence of a band of significant horizontal divergence and hence descent if the perpendicular flow component is neglected. This is not a strictly justifiable assumption in the curved flow, but its validity could be tested using the Doppler radar velocity observations. Along with the significant reflectivity values, such descent is qualitatively similar to the feature
that CF highlighted within a rear cloud edge in FRONTS 87 observations, though in
the present situation the presence of a saturated rearward circulation centred at 150 km,
700 hPa, may further contribute to forcing of local divergence.

It is helpful to consider how the gap between the cloud bands might have formed
as a consequence of sublimation-induced cooling occurring from the cloud layer over
a long period. Assume that the primary driving of such structure is latent heating and
hence ascent, which will be slantwise in the presence of thermal-wind shear. This will
form ice precipitation, which may then fall into regions which are not forced to ascend.
By the mechanism discussed in section 2 this will cause sublimation, cooling and hence
descent which, in statically stable conditions, must also be slantwise. Since, as we noted
in section 2, air at higher temperatures (up to 0 °C) can take up more vapour, and hence
cool more, we anticipate that the effect will be strongest when the precipitation and
associated descent penetrates to the upper part of the melting layer and the region above
it. This mechanism will thus cause a collapse of the cloud as precipitation descends at
the rear edge, so that hydrostatically the source of the upper cloud is cut off and a clear
band is produced behind it, while the leading edge of the sloping descent will be associated
with convergence and hence secondary ascent. Sonde 6, then, effectively corresponds to
the rear edge of the main cloud mass, as the reflectivity pattern in Fig. 5(f) suggests.

The mechanism we have just described seems to be unavoidably an evolutionary
and dissipative one in realistic systems. A steady state might be envisaged in the form
of a mode progressing towards warm air in which air ascends at the leading edge to
drive the descending circulation towards its rear. Since, however, the precipitation falls
downstream of the ascent in wind shear, the descending part of the circulation does not
precisely underlie the ascending part, and so a mesoscale break-up in structure is to be
expected. The representation of such coherent mesoscale internal structure appears to
represent a significant problem to NWP because its scales are close to typical mesoscale
grid mesh sizes both horizontally and vertically, but below those of coarser models.

The remaining dropsonde runs show much other frontal structure, particularly at
lower to mid-levels, and studies of these observations are proceeding. None, however,
shows a mid-level feature comparable to the dramatic shallow forward circulation that
we emphasise here. In part this is due to the fact that only this run occurs in the
warm-frontal part of the system. However, we note also that air parcels in this part of
the system, where the along-system flow speed is close to the wave’s velocity, have
suffered the effects of the most intense ice precipitation for the longest periods and hence
would be expected to show the greatest localized effects of sublimation, as suggested by
Clough in GEWEX (1996).

5. MODEL STUDIES

(a) Model experiments

Model experiments were carried out with the original LAM version of the Unified
Model (Cullen 1993) and high-resolution mesoscale versions initialized from the LAM
at 2100 UTC 16 February 1997. This time was chosen because the LAM forecast from
1800 UTC was judged to be the best available integration of the event, and beginning
at 2100 UTC gave three hours for relaxation of initial transients in the LAM, sufficient
time for the system to enter the mesoscale domain, and a reasonable settling time in the
mesoscale experiments themselves. The model details are set out in Table 1.

The model experiments were carried out as follows. The LAM integration A
(Table 1) was equivalent to the 1800 UTC 16 February operational LAM forecast
run, and carried out using the same model configuration though with updated physics
TABLE 1. MODEL DETAILS

<table>
<thead>
<tr>
<th>Run</th>
<th>Grid-length km</th>
<th>Number of levels</th>
<th>Precipitation scheme</th>
<th>Special features</th>
</tr>
</thead>
<tbody>
<tr>
<td>A</td>
<td>50</td>
<td>19</td>
<td>Single phase</td>
<td></td>
</tr>
<tr>
<td>B</td>
<td>12</td>
<td>45</td>
<td>Mixed phase</td>
<td></td>
</tr>
<tr>
<td>C</td>
<td>12</td>
<td>45</td>
<td>Mixed phase</td>
<td>No sublimation cooling</td>
</tr>
</tbody>
</table>

routines. It was carried out using the original operational data (i.e. a start dump from the previous LAM integration, boundary conditions from the global model and observation files), and run out to 1500 UTC 17 February. This integration used the cloud scheme then in use in the model, which treated water as vapour or liquid prognostically and diagnosed ice. Later experiments with the LAM using the new mixed-phase cloud and precipitation scheme produced very similar results in regard to the features of interest here.

The mesoscale integrations B and C were carried out on a 229 × 158 point 12 km grid, corresponding to the resolution in The Met. Office’s operational mesoscale version of the Unified Model. The large horizontal domain was determined by the need to keep the fast-moving system within the domain for the times of interest. Forty-five vertical levels were used, most of which were the same as are used in the operational 38-level mesoscale model, but with extra levels in mid troposphere (the region of primary interest in this study) to bring the vertical resolution in this region to about 30 hPa at 700 hPa. The integrations were initialized from the LAM experiment A at 2100 UTC 16 February 1997, and the boundary conditions were also obtained from this integration. All experiments were run as far as 1200 UTC 17 February, the times being chosen so that the surface feature rapidly develops from a trough to a closed low within the integration period. The two mesoscale integrations used a mixed-phase cloud scheme (Wilson and Ballard 1999) which treats liquid and ice water as prognostic variables and attempts to model explicitly the various phase transitions and interactions, including sublimation. For all three integrations a number of key dynamical variables were output on model levels and then processed with a diagnostic package to interpolate to 25 hPa pressure levels and generate derived quantities.

Experiment C was identical to experiment B except that the cooling due to the sublimation of ice was set to a very small value, essentially zero, so that the cooling due to sublimation was effectively omitted. This was done because our observational analysis, and the earlier work of CF, pointed very strongly to this cooling as the most likely driving mechanism for the descending circulation highlighted earlier, which is the essential test hypothesis of this work.

It should be noted that the precipitation scheme only carries one ice-phase variable, which is given a mean terminal velocity so no suspended ice cloud is represented.

(b) Results

Figure 6 illustrates the overall performance of the three model integrations by comparing model- deduced cloud-top fields with infrared satellite imagery. The model integrations all agree with the trajectory of the low pressure centre up to 1200 UTC, though satellite imagery and sounding data suggest that the model system may be around 150 km ahead of the observed system. It is evident also that all models, including the LAM (Fig. 6(a)), have captured the overall depth of the system. The low altitude simulated cloud tops south of the low centre suggest that the model has excessive descent
Figure 6. Model simulations of satellite infrared imagery for model experiments: (a) A, (b) B, and (c) C, at 0600 UTC 17 February 1997, compared with (d) that observed by Meteosat. The simulations are generated by plotting the temperature at the highest level at which the model cloud ice water content exceeds $5 \times 10^{-6}$ g g$^{-1}$, or else surface temperature if no ice cloud is present. (This represents frontal, though not convective, cloud reasonably well and is not critically dependent upon the threshold value adopted.) (a), (b) and (c) are overlaid with simultaneous model mean-sea-level pressure contours (hPa), while (d) shows the central pressure minimum deduced from surface, sounding and National Oceanic and Atmospheric Administration P3 aircraft observations. Positions marked A and B on (b) and (d) indicate the locations of a double cloud-head structure. The key to (a), (b) and (c) is given below left, whilst the more highly resolved (d) has its key below right. See text and Table 1 for details of experiments.
tending to form too strong a dry intrusion, a view which is confirmed by the humidity pattern in dropsonde cross-sections for the middle runs (not illustrated).

The two mesoscale simulations are both characterized by smaller-scale variations than the LAM, and they differ in the structure of the cloud head and the associated mesoscale surface pressure features. The full simulation (experiment B, Fig. 6(b)) shows rather more mesoscale features, including the differentiation of two cloud heads (Forbes et al. 2000b) with a near-break in the cloud top between—a feature which is missing from experiment C, and a tighter frontal trough closely followed by pronounced ridging. These symptoms indicate that sublimation cooling plays an important role in determining the mesoscale structure, and may even be responsible for the evolution of distinct cloud heads. The similarity of forecast central pressures seems to follow essentially the insensitive behaviour we predicted in section 2 on the basis of latent-heat exchanges due to condensation and sublimation.

We next consider the structure of simulated cross-sections compared to the observations discussed in section 3. The cross-sections were taken along the line AB marked in Fig. 10. This corresponds to essentially the same system-relative location as the observations, 150 km ahead of the observed system at the same time as in our earlier discussion. It will be evident in our later discussion that the potential-vorticity pattern in this region is quite a sensitive measure of the diabatic processes, particularly the occurrence of sublimation, and thus the position of the key circulation feature is quite precisely determined, to within about 100–200 km.

(i) **LAM experiment A.** Figure 7 shows a set of cross-sections from the LAM experiment A in a comparable location relative to the system as the observations shown in Fig. 5. Despite the model’s good reproduction of the pressure field, the vertical structure and mesoscale variations are greatly lacking compared with the observations. In Fig. 7(a) the overall range of $\theta_w$ values is reasonably reproduced, but there are only weak variations in gradients compared with those observed and no upper front is evident. The dry static stability, $N^2$, in Fig. 7(b) also shows much less variation than observed (Fig. 5(b)); a region of increased values exists above 550 hPa, above a local minimum, but the clearly differentiated layering structure of the observations is lacking.

Thus there is a minimal reproduction of the observed tendency to stratification in this model, but the ranges and length-scales of variation are quite incorrect, particularly in the vertical. A very deficient distribution of the incidence of cloud and convection might be expected to result from this lack of adequate information on the air mass stability structure. The moist static stability from Fig. 7(a) shows no convective instability other than immediately above the sea surface, whereas the full-resolution soundings show zones of weak mid-level convective instability around 600 and 700 hPa, which are shaded in Fig. 5(a).

The section-parallel flow in Fig. 7(c) also shows qualitatively a reasonable pattern, though lacking the observed maximum rearward flow at 700 hPa and strong gradient variations compared with Fig. 5(c). Most of the region is at or above saturation with respect to ice (Fig. 7(d)) without the observed mesoscale relative-humidity structure of Fig. 5(d). Although this model includes parametrization of sublimation and the other moist processes operating at the mesoscale, it appears that the vertical resolution is simply insufficient to capture the important observed signatures. We thus infer that the mesoscale processes, including resolved frontogenesis, which determine vertical structure are not adequately described by the LAM. Comparing the momentum cross-sections in Fig. 7(e) with those of Fig. 5(e), the contrast between resolvable strong and
weak gradients marking zones of modest and low symmetric stability is suggested but not well captured.

Overall we conclude that, despite the reasonable synoptic performance, this integration is lacking in the mesoscale structure on scales of interest.

(ii) Mesoscale model experiment B. The results of experiment B are illustrated in Fig. 8. These cross-sections bear a clear structural similarity to those shown in Fig. 5,
and have well-defined mesoscale humidity variations and internal structure. Sloping layers of high and low static stability are now present (Fig. 8(b)), though they occur at a level almost 100 hPa below those observed. The levels and depths of the regions of high and low static stability relative to the upper front are in much better agreement with the observations than those in experiment A.

Collocated with the sloping layer of strong static stability is a shallow layer of increased forward descending flow (Fig. 8(c)) in the upper front. It is weaker than that observed, but qualitatively similar and in a similar location relative to the thermal gradient. The humidity pattern (Fig. 8(d)) shows some qualitative similarity to the observed humidity variations, and the region of the strong forward flow possesses a gradient with values decreasing from 100% to around 98% of saturation with respect to ice, very much in line with the predictions of CF. However, the model variations are rather smaller in amplitude than the observed humidity fluctuations and also show much smaller supersaturation with respect to ice, a feature also shared by the other runs. This suggests that the model moisture scheme may have a significant deficiency on NWP time-scales; with limited potential instability much of the moisture transport at this resolution would be expected to occur on or near to resolved scales. If it indeed represents a systematic bias it implies rather different mesoscale precipitation intensity and moisture-driven transfers in mid troposphere. Note that surface precipitation may not be dramatically affected by such an error since most of the water content in
the column occurs near the surface; nonetheless the dynamical effects highlighted by this work would be inadequately represented if the observed supersaturation and precipitation rates aloft are not achieved.

Figure 8(e) shows the momentum, which possesses a shallow slope through the upper front much like that observed qualitatively, though the height of the strong wind shear zone is too low like the other frontal characteristics. Since momentum is conserved
in two-dimensional flow, the shallow slope would be associated with a direct cross-frontal circulation, which is the case here. Thus, if our dynamical coupling interpretation holds, it seems quite possible that the deficiency in the height of the thermodynamic structure could feed into the jet and momentum structure through the cross-frontal flow.

The structure of the above modelled features corresponds both to that observed and to the Clough–Franks paradigm. It should be noted, however, that the Clough–Franks mechanism was derived in an idealized two-dimensional context, whereas the real situation, as here, inevitably has a strongly three-dimensional element.

(iii) Mesoscale model experiment C. Figure 9 presents the cross-sectional structure of the experiment C simulation, in which the hypothesized thermodynamic effect, i.e. cooling due to sublimation, has been omitted.

It is evident in both Fig. 9(a) and (b) that the strong contrast between weakly and strongly stable zones in experiment B is reduced. Also the shallow forward descending flow is much weaker in Fig. 9(c), and the amplitude of the humidity variations is reduced (Fig. 9(d)). Thus the stability and circulation symptoms that we have associated diagnostically with sublimation are, indeed, both associated with sublimation in the model, and a pronounced sensitivity of the flow results.

Figure 9(e), however, shows that the structure of momentum surfaces is much more like that of experiment B than that of experiment A, though with less intense gradients. Theoretically, the momentum surfaces are most directly associated with
the potential vorticity, or moist potential vorticity, which determines the Jacobian constraining their slopes compared with the potential-temperature distribution in two dimensions (Bennetts and Hoskins 1979). Figures 10 and 11 present horizontal and vertical sections of the dry potential-vorticity structure from the two experiments, which is qualitatively very similar to the moist potential-vorticity pattern. Figure 10 shows plan views of potential vorticity at 525 hPa from experiments B and C, with the location of
vertical cross-sections marked. Even without sublimation Fig. 10(b) shows low potential vorticity in experiment C at this level, negative in places, because the latent heating is greater at lower levels. Clearly, sublimation cooling in experiment B has a dramatic effect upon the mesoscale potential-vorticity pattern in Fig. 10(a), with a minimum value of $-2$ PVU ($1$ PVU = $10^{-6}$ m$^2$s$^{-1}$ K kg$^{-1}$) being reached. The strong section-parallel flow in the presence of sublimation, corresponding to that observed, is localized to the 100 by 200 km region between the positive and negative maxima of an intense dipole. The dipole at this level is part of a sloping potential-vorticity pattern consisting of a sheet of negative values overlying a sheet of positive values following a warm-frontal plane approximately perpendicular to AB in Fig. 10(a) (not illustrated). A shallow descending jet flows anticyclonically along the warm surface plane between the potential-vorticity sheets in run B, fuelled by the maximum in sublimation cooling immediately beneath the frontal plane. Viewed in this way the observed structure and its physical origin are much more readily placed in their system context.

Figure 11 shows that the potential vorticity in the vicinity of the shallowest slopes of momentum is near-zero in both experiments B and C. The occurrence of a significant region of negative potential vorticity implies that SI (or CSI, since the moist potential vorticity is also negative) is possible in this location. This might in principle be problematic, since the model's vertical resolution is less than that found by Persson and Warner (1991) to be necessary for consistency with the horizontal grid to properly
Figure 10. Potential vorticity at 525 hPa at 0600 UTC 17 February 1997 from experiments (a) B and (b) C; Line AB shows the location of the cross-sections, and L the low centre. Values are shown by dashed contours at intervals of 1 PVU ($10^{-6} \text{ m}^2\text{s}^{-1} \text{ K kg}^{-1}$). Values less than $-1$ PVU are lightly shaded, while values greater than $1$ PVU are heavily shaded. Solid contours are surface pressure with contour interval 2 hPa.

Figure 11. Potential vorticity from experiments (a) B and (b) C along the line AB marked in Fig. 10. Contour interval is 0.5 PVU ($10^{-6} \text{ m}^2\text{s}^{-1} \text{ K kg}^{-1}$); see key for values.
Figure 12. Vertical velocity (m s\(^{-1}\)) from experiments (a) B and (b) C along the line AB marked in Fig. 10. Contour interval is 0.02 m s\(^{-1}\).

resolve gravity waves, particularly in almost symmetrically unstable conditions. Under these conditions some distortion of the vertical velocity, and hence potential-vorticity pattern, is possible near the grid-scales, and the fastest growing CSI modes may not be adequately described. Higher resolution is not practicable at the present time, but we have carried out an integration with doubled horizontal grid spacing, and hence more acceptable aspect ratio; this resulted in smoother fields but otherwise little significant change in our results. Our above results do suggest, however, that if SI or CSI is important in the resulting mesoscale development, then sublimation cannot be neglected since it has such substantial effects upon the potential-vorticity pattern, particularly in the sensitive negative regime where resistance to descent through static stability is appreciably modified. In this situation an important factor in determining the mesoscale structure will be the quantity of precipitation available to evaporate, hence the importance of the overall reflectivity and relative-humidity distribution.

The vertical velocity cross-sections from the model experiments B and C are shown in Fig. 12. As expected from the horizontal flow, both experiments possess shallow regions of descent coincident with the forward flow, and the descent in experiment B is greater by a factor of around two. Figure 13 shows the cross-section of moist static stability, \(N^2\), from experiment B. The effective static stability in this moist case, is smaller than the dry static stability which applies in experiment C (Fig. 9(b)) in the forward flow by a factor of around 1.7. This behaviour is also consistent with
moist adiabatic descent in precipitation, noting that the forcing of descent has been increased slightly because of the lower negative potential-vorticity values (i.e. more anticyclonic flow) in run B. With this allowance, our results confirm in all respects that CF's prediction of moist slantwise descent provides a consistent account of both the atmospheric and model results.

6. A SUBLIMATION HYPOTHESIS FOR WEATHER SYSTEM STRUCTURE

It is possible to generalize from the above results. Since ice precipitation is usually a concomitant of ascent in frontal waves, sublimation potentially provides a feedback by which any forcing of ascent also forces descent, and this descent must occur on the mesoscale because of the short length- and time-scales of precipitation itself.

By this mechanism ice precipitation directly provides downward transfer of dynamical influence by its presence, and possibly hydrostatic adjustment to the cooling will also imply some upward influence. Since precipitation falls primarily from the ascending downstream part of a wave, it may be a means of transferring the effects of forcing through mesoscale distances downstream in the warm-sector part of a system, and also across-stream if it operates in conjunction with slantwise ascent or descent. Precipitation does also fall into air parcels moving upstream beneath the critical level, but its effects seem likely to be less distinct since these parcels only transiently experience precipitation as they emerge beneath the descending drier airmass in the rear of the wave without upper and mid-level cloud.

The above processes thus provide a precisely defined means by which ascent and hence precipitation, though not necessarily mesoscale in origin, will have mesoscale substructure associated with them, without any need to hypothesize the action of a mesoscale dynamical instability such as CSI. The scales of this substructure are determined not by mesoscale dynamics in the accepted sense but by the microphysics of the precipitation processes, particularly sublimation, and their interaction with system kinematics as discussed below. (There is also much evidence for the important role of melting in organizing mesoscale structure of strongly precipitating frontal waves, as demonstrated by Stewart and co-workers (e.g. Szeto and Stewart 1997).)

We therefore make a sublimation hypothesis, that sublimation is normally associated with the development of broad rain bands in middle-latitude frontal systems. This process model has some features to commend it. In particular it is distinct from current conceptual models in being process-specific, conservative (in that it accounts
for the coupled changes in all conserved variables including moisture), and indicates quite specifically the circumstances in which sublimation will occur, the limits on its amplitude and the necessary observables for verification. These observables include the conventional dynamical variables and the distribution and properties of ice precipitation observable by aircraft or radar.

7. Conclusions

In this study we have provided clear observational and model evidence of the importance of sublimation in producing mesoscale dynamical structure; to that extent it follows condensation in the hierarchy of dynamical processes of frontal waves. The structure coincides with that predicted by Clough and Franks (1991) from microphysical calculations and dynamical arguments; in particular it confirms that sublimation can support very nearly moist-adiabatic slantwise descent, thus providing an enabling machinery for the development of mesoscale precipitation bands. This hypothesis was confirmed by omitting sublimation cooling from the model, which led to a substantial weakening of the characteristic symptoms in static stability and circulation. We suggest that the resulting effect be referred to as Sublimation Enhanced Descent or SED.

The close correlation between the observed and modelled section-parallel circulation and the moisture distribution strongly suggests that microphysical characteristics are a major factor affecting mesoscale dynamical structure within frontal waves. Since current cloud/precipitation parametrizations have been little optimized with regard to microphysical characteristics and the ambient conditions, they may prove to be a key element in improving NWP using high-quality data such as those available from FASTEX.

Our results have several important implications for our understanding of, and the quantitative forecasting of, frontal waves in terms of microphysical processes and their role in system dynamics.

- In view of the active role of ice precipitation in driving mesoscale circulations we suggest that precipitation rate, typically observed as reflectivity, should properly be viewed as a strongly coupled dynamical variable of the system. It should be measured where possible along with the other thermodynamic elements, because of its importance as a vertical-coupling agent operating on short time-scales compared with those of balanced flows.

- Our demonstration that frontal circulation and static-stability structure are sensitive to sublimation provides important verifiable links between observed thermodynamic and dynamical structure, and also between the fundamental microphysical processes and precipitation structure. Since CF have shown that the dynamical effects of sublimation are limited by the concentration of precipitation, a sensitivity of dynamics to the whole vertical profile of precipitation and microphysical properties such as terminal velocity is evident.

- Our results suggest that the system studied behaves approximately linearly in the imposed forcing, in that the vertical velocity increased by about a factor of two while the effective static stability, $N^2$, decreased by 1.7, using the appropriate static stability for the conditions, which can be moist adiabatic through depth-scales for which sublimation of snow is sufficient to support it.

- In our model simulations SI arose as a local outcome of diabatic processes, rather than itself driving the system's overall behaviour on a larger scale. This is encouraging from the point of view of understanding weather system development, since stable forcing/response systems are much more readily understood and quantitatively forecast
than the growth of inherently unstable modes. Negative potential vorticity was found to be present without sublimation, but was appreciably enhanced in the presence of sublimation cooling. Thus, the forecasting problem in this situation appears to be one of internal complexity of well-understood processes and possibly model resolution, rather than unpredictability associated with CSI or other mesoscale dynamical instability.

- The model results suggest that atmospheric structure in the region observed should be studied closely using radar and aircraft observations, in order to refine our understanding of the atmosphere’s response to such forcing in conditions potentially unstable to SI or CSI. It is interesting to note that SI, if it indeed occurred in the atmosphere, is here in the warm-frontal region of the warm sector in mid troposphere in the anticyclonic outflow from the main latent-heating region, a location which has attracted very little attention in observational studies. However, at this location interactions between cloud physics and system dynamics are likely to be particularly important because of the long periods during which mid-level air parcels may be diabatically modified within the ice precipitation envelope. Since the symptoms associated with warm fronts typically extend into mid-troposphere, these processes need to be addressed in any proper quantitative understanding of warm frontogenesis.

- In discussing banding in the corresponding satellite imagery, and its evolution explicitly in terms of ice precipitation and its sublimation, we adopted a direct physically based approach to satellite image interpretation. From these considerations we suggested that observed divergence fields may be modified by the precipitation processes and perhaps provide a forcing for deep bands of mesoscale divergence noted by Thorpe and Clough (1991) in FRONTS 87 observations. This possibility could be explored much more precisely in Doppler radar observations. We recommend that this viewpoint be considered in interpreting weather system structure and evolution more generally, since our inferences refer to symptoms observed daily in real time in satellite infrared and other imagery.

There are however some important issues that we have not addressed here, some of which are presently being studied.

Our integrations, for practical reasons of domain size and resolution, have covered only a 9–15 hour period. Attempts are being made to initialize from earlier times, to assess whether a better simulation will result from better resolution of the sublimation, and to assess the overall role of sublimation in determining localized frontal structure within the system.

The situation described by Harris (1977) has not arisen in an important way in the present case. Shallow layers of static instability were evident in some high-resolution sonde data for this and other cases, but the present model simulations do not approach static instability aloft. The possibility that sublimation of ice precipitation may induce layers of instability remains to be investigated more broadly. Its occurrence and distribution will depend upon the presence of air parcels which persistently encounter precipitation by travelling at close to the system’s velocity, a situation which must occur most often in the warm sector and warm-frontal region aloft. Both are regions where banded precipitation frequently occurs at the surface, though very few studies have been made of the associated mesoscale structure aloft.

In demonstrating that the feature studied is amplified by sublimation-enhanced feedback, we transfer attention to the forcing process. It is possible, for example, to envisage that the simplest baroclinic system, a moist Eady wave forced only by boundary temperature perturbations, may generate a wealth of mesoscale internal structure in which sublimation has operated upon any forcing for descent to amplify it within the
precipitating regions. As in our preceding discussion, the size of these effects will clearly be proportional to the strength of the vertical motion, but also their distribution will depend upon the interaction between sublimation and system kinematics. The location of the structure studied above is consistent with this expectation, which was originally suggested by Clough on the basis of earlier observations of mid-tropospheric static-stability structure (GEWEX 1996).

When we focus attention upon the kinematic properties of the system we encounter an interesting corollary. Bretherton (1966) highlighted the importance of critical-layer instability in the theoretical analysis of baroclinic instability. If, as envisaged above, a simple Eady wave can develop mesoscale internal structure via precipitation and sublimation, that structure will tend to form in the vicinity of the wave's critical surface in mid troposphere where the vertical motion is greatest and residence time longest. Thus, the shape of the critical surface is likely to determine the resulting mesoscale structure, as has been found in studies of larger-scale diabatic frontogenesis (Hoskins and West 1979; Davies et al. 1991).

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