Dynamics and Organisation of Precipitation Bands in the Midlatitudes

A thesis submitted to the University of Manchester for the degree of Doctor of Philosophy (PhD) in the Faculty of Engineering and Physical Sciences

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Abstract


This thesis was funded by the Natural Environmental Research Council (NERC) as part of the Diabatic Influences on Mesoscale Structures in Extratropical Storms (DIAMET) project. The thesis is presented in alternative format, meaning that the results of the thesis take the form of three journal articles, each telling a distinct story within the subject matter, but collectively highlighting the sensitivity of bands to frictional and diabatic processes.

Paper 1 is an idealised-modelling study with the Weather Research and Forecasting (WRF) model, in which moist baroclinic waves are simulated from an initial zonally uniform midlatitude jet on an $f$-plane at 20-km grid spacing, and the sensitivity of the resulting precipitation bands is explored. Paper 2 employs further WRF idealised-baroclinic-wave simulations and takes a simulation from Paper 1, after the cold front has formed, as the initial condition. A nested domain at 4-km grid spacing is inserted when this simulation is re-initialised to investigate the sensitivity of finer-scale precipitation cores along the surface cold front. In both Papers 1 and 2, friction and latent-heat release enhance multiple banding at the two distinct horizontal scales, while surface fluxes hinder multiple banding.

Paper 3 studies postfrontal snowbands over the English Channel and Irish Sea during extreme cold-air outbreaks in the winters of 2009–10 and 2010–11, via a climatology of precipitation-radar, sounding, and SST data, and real-data WRF sensitivity simulations of one such band over the English Channel. The observational and modelling results show that strong winds and large differential heat fluxes between land and sea were necessary to generate banded precipitation. Coastal orography and the land–sea frictional contrast aided the morphology of bands, but banded precipitation did still form in the absence of these influences in the sensitivity simulations.

These three studies and the thesis as a whole highlight the role of frictional and diabatic processes in modifying various types of precipitation bands within baroclinic waves, and in generating bands that would otherwise not exist.
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1 Introduction

Heavy precipitation in the midlatitudes is often organised into bands. A band is a precipitation feature, which to a good approximation can be taken to be two-dimensional in the $n$–$z$ plane, where the $n$-axis is normal to the band’s major axis, and the band’s length is much greater than its width. Figure 1 exhibits two distinct categories of bands that form in the midlatitudes, as observed by precipitation radar (see chapter 2.1 for a description of precipitation radar). In Fig. 1a, the rainfall along a surface cold front has divided into multiple precipitation cores of clockwise orientation, relative to the cold front, with gaps of lighter or no precipitation between cores. In Fig. 1b, a single snowband lies over an elongated body of water, parallel to that body of water’s major axis.

This thesis is concerned with an improved understanding of the distribution of precipitation bands within midlatitude weather systems, as well as an improved understanding of these two specific band types. The bands in Fig. 1a formed during the passage of an extratropical cyclone, whereas the band in Fig. 1b formed in the absence of an extratropical cyclone, during a cold-air outbreak. Extratropical cyclones and cold-air outbreaks are both products of baroclinic waves, whose theory underpins this thesis.

1.1 Baroclinic waves

A synoptic-scale depression can be viewed as a positive or negative potential-vorticity (PV) anomaly in the case of a low- and high-pressure centre, respectively (Hoskins et al. 1985). Ertel PV (Ertel 1942) is proportional to the vorticity of the flow relative to an isentropic surface:

$$ q = \frac{1}{\rho} \zeta \cdot \nabla \theta $$

where:

$$ \zeta = \begin{pmatrix} 0 \\ 0 \\ f \end{pmatrix} + \nabla \times \mathbf{u} $$

is the 3-D absolute-vorticity vector, $\rho$ is density, $\theta$ is potential temperature, $f$ is the Coriolis parameter, and $\mathbf{u}$ is the 3-D wind vector.

The use of PV as a diagnostic variable is based on two principles, which aid the analysis of weather systems. The first principle is that, if PV is defined on isentropic surfaces and the mass under each isentropic surface is known, together with $\theta$ at the
Figure 1: Met Office precipitation-radar composites at 1-km grid spacing, expressed as mm h\(^{-1}\), showing two different types of precipitation bands over the British Isles: (a) Precipitation cores along a cold front at 1600 UTC 29 Nov 2011; (b) A quasi-stationary snowband over the English Channel during a cold-air outbreak at 0000 UTC 1 Dec 2010.
lower boundary, then, given an appropriate balance condition, the PV distribution can be inverted, to give all dynamic variables (Hoskins et al. 1985). The second principle is that, following air parcels, PV is conserved in the absence of friction and diabatic effects (Rossby 1940; Ertel 1942). Therefore, when some disturbance occurs in the atmosphere, the diagnostic of a PV anomaly allows air parcels to be traced approximately to their origin.

1.1.1 Baroclinic instability

Individually, a PV anomaly does not amplify, but if coupled with another sufficiently close for the advection of air parcels generated by one to be felt at the other, then the two PV anomalies may interact. If the atmosphere is suitably unstable, described below, then their interaction can lead to phase locking and mutual amplification.

PV anomalies formed on positive and negative PV gradients propagate left and right, respectively, relative to the background flow (Charney and Stern 1962). Charney and Stern (1962) showed that, for an instability to exist, the meridional PV gradient, $q_y$, must be of opposite signs at the locations of two separate anomalies. Fjortoft (1951) showed that, if:

$$ u_{q_y > 0} > u_{q_y < 0} $$

where $u_{q_y > 0}$ and $u_{q_y < 0}$ are the zonal wind speed at the locations of the anomalies where $q_y$ is positive and negative, respectively, then the wind shear forces the two anomalies into a mutually amplifying configuration. The situation where two PV anomalies satisfying (2) are formed at different latitudes and interact across the $y$-axis is known as barotropic instability. Alternatively, two anomalies may be formed at the same latitude but at different altitudes and so interact in the vertical (across the $z$-axis). This situation is known as baroclinic instability (Fig. 2).

These large-scale PV waves are known as Rossby waves (Rossby 1940). Near a lower boundary, warm and cold anomalies induce cyclonic and anti-cyclonic circulations, respectively (Bretherton 1966), and so act as positive and negative PV anomalies. Therefore, the lower Rossby wave in Fig. 2 can form above a lower boundary if warm air to the south is displaced to the north. Conversely, a warm anomaly near an upper boundary acts as a negative PV anomaly. Therefore, the upper Rossby wave in Fig. 2 can form beneath an upper boundary when warm air is displaced north. In the Earth’s atmosphere, the tropopause acts as this upper boundary. However, unlike the surface, the tropopause is not rigid and so vertical
motion is generated across the tropopause to balance the circulations associated with the PV anomaly. The resulting descent of statically stable air from the stratosphere into the troposphere is known as the tropopause fold (Fig. 3).

Thus, in the Earth’s atmosphere, baroclinic instability exists via the phase-locking of two Rossby waves, one near the surface and one near the tropopause. From (2), in order for mutually amplifying Rossby waves to exist, a large-scale vertical shear of the zonal wind is required. This wind shear exists because of the large meridional gradient of potential temperature that arises from the differential of received solar radiation between the equator and poles. This $\theta$-gradient results in vertical wind shear, due to approximate thermal wind balance:

$$ f \frac{\partial u}{\partial z} = -\frac{g}{\theta_0} \frac{\partial \theta}{\partial y} \quad (3) $$
where $g$ is acceleration due to gravity and $\theta_0$ is a reference value for $\theta$. This vertical wind shear implies strong eastward winds in the upper troposphere, known as a jet stream. Extratropical cyclones form as low-pressure perturbations to the polar jet (Fig. 4), which amplify over time if the jet is baroclinically unstable and gain their energy from the high winds in the jet. The full perturbation to the jet, which includes the high- and low-pressure anomalies, is known as a baroclinic wave.

1.1.2 Fronts

A front is a boundary separating air masses of different temperature, formed by the tightening of a horizontal thermal gradient $\left(\frac{\partial \nabla \theta}{\partial t} > 0\right)$ by the synoptic-scale flow. The formation of fronts at the surface by the cyclonic advection of air parcels induced by a synoptic-scale low centre is depicted in Fig. 5. As shown in the diagram, the isentropes relative to a cyclone become organised into two fronts. The cold front to the south marks the boundary of the cold air that is advected around the low centre from the north. The warm front to the north marks the boundary of the warm air that is advected around the low centre from the south.
The Norwegian cyclone model

The Norwegian cyclone model (Bjerknes 1919; Bjerknes and Solberg 1922; Fig. 6a) remains a template against which extratropical cyclones are described. The surface cold front is oriented meridionally and marks the boundary of the cold air that is advected around the low centre from the north to the west. The surface warm front is oriented zonally and marks the boundary of the warm air that is advected around the low centre from the south to the east. The warm air between the cold and warm fronts is known as the warm sector. If there is an upper-level PV anomaly with which this surface anomaly can mutually amplify, then the low deepens and the fronts intensify. The Norwegian model depicts the cold front eventually catching up to the warm front, but Schultz and Vaughan (2011) argued that, instead, the isotherms collectively wrap around the low centre. In the Norwegian model, the warm-sector air forms an occlusion marked by the occluded front, separating two cold air masses at the surface. The occluded front extends from the apex of the warm sector poleward towards the low centre.
Figure 5: The strengthening of temperature gradients at the surface, due to advection of air parcels associated with a low-pressure centre, resulting in the warm front to the north and cold front to the south. Solid contours are isentropes, weak contours are isobars, and dashes are axes of dilatation of the horizontal wind. Taken from Schultz et al. (1998).

The Shapiro–Keyser cyclone model

In contrast, Shapiro and Keyser (1990) observed that the structure and evolution of certain midlatitude cyclones differ fundamentally from that of the Norwegian cyclone model and developed the Shapiro–Keyser cyclone model (Fig. 6b). In this conceptual model, the cold and warm fronts initially form, as in the Norwegian model, but move perpendicular to one another, forming a T-bone shape. The poleward end of the cold front begins to weaken, forming a well-defined gap between the cold and warm fronts, known as the frontal fracture. The warm front then wraps around the low and forms a bent-back front. As this happens, the cold air advected around the bent-back front encircles a warm pool of air, known as the warm seclusion. Schultz and Vaughan (2011) argue that the bent-back front of the Shapiro–Keyser model is in fact an occluded front, which wraps around the cyclone centre, as in the Norwegian model, and that the warm seclusion is in fact the warm-sector air that has become detached from its source.

This interpretation suggests that the Norwegian and Shapiro–Keyser cyclone models are two ends of a spectrum, rather than fundamentally different cyclone types. In the Shapiro–Keyser model, the fronts collectively remain zonally elongated and the warm front is dominant, whereas, in the Norwegian model, the fronts become more meridionally elongated and the cold front becomes dominant.
1.1.3 Anticyclonic blocking

The low-pressure anomaly within a baroclinic wave is generally associated with the most unsettled weather because of the tendency for ascent and hence cloud formation associated with low pressure, and because the low centre produces the warm and cold fronts. The high-pressure anomaly within a baroclinic wave is generally associated with the most settled weather because the synoptic-scale descent associated with a high-pressure system inhibits cloud formation. However, high-pressure systems can remain quasi-stationary for a number of days, and even weeks, leading to what are known as blocking anticyclones. Blocking anticyclones are the product of very long, high-amplitude baroclinic waves, which result from the nonlinear interaction of long and short waves (Carlson 1991, p. 335). When blocking occurs, the polar jet is deflected around the anticyclone and prevented from following its natural west–east course. Some location beneath the anticyclone may then become subject to airflow from the same direction over a long period of time, depending on where the high-pressure centre is positioned, relative to that location. For example, in September 2011, an anticyclone became established over central Europe, leading to prolonged southerly flow over the UK (Fig. 7a) and hence unseasonably warm weather. Conversely, in the winters of 2009–10 and 2010–11, a succession of anticyclones became established over the North Atlantic, leading to prolonged northerly flow over the UK (Fig. 7b) and hence bitterly cold weather.
Figure 7: Met Office sea-level pressure analyses during anticyclonic-blocking episodes where the UK was subject to (a) a warm-air outbreak at 00 UTC 30 Sep 2011; (b) a cold-air outbreak at 00 UTC 17 Dec 2010. Pressure contoured every 4 hPa. Frontal notation is standard. Solid lines represent troughs.
Although blocking episodes may be considered distinct weather regimes to extratropical cyclones, in Fig. 7a, the southerly flow over the UK is associated with the warm sector of a cyclone over the North Atlantic. In Fig. 7b, the northerly flow over the UK comes behind the cold front of a cyclone over the North Sea. These examples illustrate that blocking episodes are fundamentally linked to extratropical cyclones, and that the weather experienced during prolonged warm- and cold-air outbreaks are extreme instances of warm-sector and postfrontal weather, respectively.

1.1.4 Baroclinic-wave models

The theory of baroclinic waves allows extratropical cyclones to be simulated in idealised baroclinic-wave models, so that the weather associated with cyclones and factors affecting cyclones may be investigated without the complications of the real atmosphere. A baroclinic-wave model takes an idealised baroclinically unstable jet as its initial condition, and applies a finite-amplitude perturbation, which, over a number of days, evolves into an extratropical cyclone with associated cold and warm fronts (Fig. 8).

Eady model

The simplest baroclinic-wave model is the Eady model (Eady 1949), in which three major simplifications are made to baroclinic waves in the Earth’s atmosphere. First, the lower and upper boundaries are taken to be fixed. For the lower boundary, this is realistic, because the Earth’s surface is fixed in space, but taking the upper boundary to be fixed prevents the simulation of the tropopause fold. Second, the Eady model is defined on an $f$-plane, meaning that the Coriolis parameter is invariant, unlike the real atmosphere, in which $\frac{df}{dy} > 0$ (in the Northern Hemisphere). Finally, there is no horizontal shear in the initial jet, only vertical shear. Despite these simplifications, the Eady model provides the necessary ingredients to simulate a baroclinic wave, as depicted in Fig. 2. However, as illustrated below, more complex initial and boundary conditions are required to investigate the variability of cyclones in the real atmosphere.

Sensitivity to initial and boundary conditions

In the real atmosphere, the initial state of the atmosphere against which any cyclone develops is far more complex than a uniform jet in thermal wind balance. Thus, many studies have edited the initial conditions for idealised baroclinic waves
Figure 8: Evolution of an idealised baroclinic wave in a periodic domain, simulated by the Weather Research and Forecasting (WRF) model (the NOFRIC simulation in Paper 1 of this thesis), showing surface potential temperature (colours every 5 K) and pressure (contours every 4 hPa, low and high centres marked by L and H). Domain re-centred in each panel around temperature anomaly.
to investigate the sensitivity to the initial conditions and attempt to explain why different cyclones are structurally different to one another, for example, why certain cyclones adhere to the Norwegian cyclone model and others more closely to the Shapiro–Keyser model. Davies et al. (1991) found that the addition of cyclonic and anticyclonic shear across the initial jet (which is absent in the Eady model) results in the strongest winds occurring to the south and north, respectively, favouring the warm and cold fronts, respectively (Fig. 9). Similarly, Thorncroft et al. (1993) found initial jets with no shear and cyclonic shear to result in meridionally and zonally elongated thermal distributions, leading to dominant cold and warm fronts, respectively.

Schultz et al. (1998) argued that the effect of cyclonic versus anticyclonic shear is more accurately described by the contrast between confluence and diffuence in the initial jet. A cyclone forming against background confluence leads to a zonally elongated vortex, leading to a stronger warm front, whereas a cyclone forming against background diffuence leads to a meridionally elongated vortex, leading to a stronger cold front.

The lower boundary condition may also profoundly influence cyclone structure. Hines and Mechoso (1993) found that warm advection is much more sensitive than cold advection to surface friction. For a frictionless lower boundary, strong warm
advection develops so that the warm front wraps around the low centre, preventing the formation of a distinctive cold front and leading to a Shapiro–Keyser-type cyclone. As friction increases between simulations, warm advection is greatly reduced, while cold advection is relatively unaffected, leading to a stronger cold front, relative to the warm front, and a more Norwegian-type cyclone. Rotunno et al. (1998) produced similar results and argued that the difference in sensitivity to friction between warm and cold advection is because the warm front forms near the low centre, where Ekman pumping induced by surface drag is strong, while the cold front forms well away from the low centre, where Ekman pumping is weaker. Ekman pumping is the mechanism by which boundary-layer momentum fluxes are communicated to the free atmosphere by a forced secondary circulation (Holton 2004, p. 132). Adamson et al. (2006) showed that friction in fact reduces the deepening of cyclones via PV anomalies generated at the surface, partly by Ekman pumping and partly by baroclinic processes, which are transported around the cyclone by the conveyor belts.

When surface sensible heat fluxes are included in baroclinic-wave simulations, particularly over water, the displaced warm and cold air flows over a surface that is cooler and warmer than the air aloft. The resulting negative surface heat flux where there is warm advection and positive surface heat flux where there is cold advection reduce the temperature gradient at both the warm and cold fronts (e.g., Mansfield 1974; Sinclair et al. 2010; Boutle et al. 2010). Therefore, when the lower boundary is warmer (e.g., for cyclones forming further equatorward), the positive heat flux in the cold air is increased, which may weaken the cold front, but the negative heat flux in the warm air is reduced, which may enhance the warm front. Correspondingly, when the lower boundary is cooler (e.g., for cyclones forming further poleward), the cold front may be enhanced and the warm front reduced. To this effect, Nuss and Anthes (1987) found that, for idealised baroclinic waves forming above a sea-surface-temperature (SST) distribution that is in phase with the lower-tropospheric temperature distribution, the low-level baroclinicity is increased, so that the surface fronts are enhanced and the cyclone deepens more rapidly.

The above studies collectively illustrate that the large-scale structure of baroclinic waves is strongly influenced by the initial state of the atmosphere when the cyclone forms, and by the frictional and thermal properties of the lower boundary. Therefore, finer-scale features within baroclinic waves, specifically precipitation bands, may also be highly sensitive to the initial and boundary conditions.
1.2 Banded precipitation

Heavy precipitation within extratropical cyclones often takes the form of bands (e.g., Browning and Harrold 1969; Browning et al. 1973; Houze et al. 1976; Hobbs 1978; Parsons and Hobbs 1983; Browning et al. 1997; Browning 2005). Houze and Hobbs (1982) produced a conceptual model of where different types of bands have been observed to form relative to the parent cyclone, showing warm-frontal, warm-sector, wide cold-frontal, narrow cold-frontal, prefrontal-cold-surge, and postfrontal bands (Fig. 10). This conceptual model illustrates that the bands are generally aligned parallel to one of the cyclone’s fronts, motivating the question of how airstreams within cyclones result in front-relative motion and, thus, the generation of bands.

1.2.1 Conveyor belts

Many of these bands arise from airstreams within extratropical cyclones called conveyor belts (e.g., Carlson 1980). The dominant flow in producing precipitation in midlatitude cyclones is the warm conveyor belt (WCB). The ascent of air in the warm sector over the cooler air around it brings more warm and moist air from low latitudes and low altitude to be drawn poleward and upward. As the air cools (due to both the poleward and upward motion), the water vapour condenses into a trail of cloud, as in Fig. 11.

In the event of a sharp cold front, some of the WCB air that rises poleward through the warm sector undergoes rearward ascent over the cold front behind it (Browning 1986). Due to the sharp contrast between the warm air in the WCB and the cold air immediately behind it, the air in the WCB is lifted abruptly in a narrow updraft to 2 or 3 km above the surface cold front, resulting in the narrow cold-frontal rainband. Behind the narrow band, the warm air subsequently rises in a slantwise fashion up the cold front, producing the wide cold-frontal bands. This situation of upright, followed by slantwise rearward ascent, is depicted in Fig. 12 and describes an ana cold front.

In some instances, most of the WCB air continues poleward and ascends over the warm front (Browning 1986), so that there is relatively little precipitation along the cold front and a greater tendency for bands in the warm sector and along the warm front (e.g., Browning et al. 1973; Heymsfield 1979; Novak et al. 2004). In this case, dry and cold air from the upper troposphere and lower stratosphere overruns the WCB. This is known as the dry intrusion and, due to the large humidity contrast between air in the WCB and dry intrusion, leads to potential instability
(e.g., Browning et al. 1973). The leading edge of the dry intrusion advances ahead of the surface cold front, creating an upper-level front (Fig. 13), which can lead to significant precipitation ahead of the surface cold front. Such a cold front is known as a kata front. An ana front may evolve into a kata front during a cyclone’s lifecycle (Browning 1986) and a single cold front may exhibit characteristics of both ana and kata fronts at different ends of the front simultaneously (e.g., Mass and Schultz 1993; Browning and Roberts 1996).

Also observed in the system-relative flow is a cold conveyor belt (CCB) originating northeast of the cyclone centre and flowing at low levels just ahead of the surface warm front beneath the WCB (Browning 1986). The air in the CCB produces lit-
tle precipitation because it does not ascend, and much of the WCB precipitation evaporates upon falling into it. If the CCB emerges beneath the WCB, it typically encircles the low centre, as in the Norwegian and Shapiro–Keyser models. When this occurs, a comma head of cloud cover may form. The CCB is not generally associated with bands. However, when the cold air in the CCB flows over warm water, shallow convection in the boundary layer may occur, which is organized into postfrontal bands, parallel to the vertical shear of the horizontal wind, when that shear is of sufficient magnitude (e.g., Kuo 1963; Asai 1970; Miura 1986; Shirer 1986). Such a band is exhibited in Fig. 1b and explored in Paper 3 of this thesis, in the context of cold-air outbreaks (section 1.1.3).

These conceptual models of conveyor belts within cyclones motivate the question of why some cold fronts behave as ana-type and others as kata-type fronts. More generally, what factors determine front-relative motion when some synoptic-scale forcing is applied?
1.2.2 Symmetric instability

When some frontogenetical forcing is applied to an existing temperature gradient in the atmosphere, the atmosphere responds by generating ageostrophic motion in the cross-frontal plane to restore thermal wind balance (Holton 2004, p. 277). The Sawyer–Eliassen equation (Sawyer 1956; Eliassen 1962) describes the response of the atmosphere to a given frontogenetical forcing:

\[
- \left( \frac{\gamma}{\frac{\partial H}{\partial p}} \right) \frac{\partial^2 \psi}{\partial y^2} + \left( 2 \frac{\partial M_g}{\partial p} \right) \frac{\partial^2 \psi}{\partial y \partial p} - \left( \frac{\partial M_g}{\partial y} \right) \frac{\partial^2 \psi}{\partial p^2} + \left( \frac{\partial M_g \frac{d \ln \gamma}{dp}}{\partial p} \right) \frac{\partial \psi}{\partial y} = -2 J_{yp}(u_g, v_g) - \gamma \frac{\partial \dot{\theta}}{\partial y} + \frac{\partial F_x}{\partial p}
\]

where \( \psi \) is the ageostrophic streamfunction, such that:

\[
v_{ag} = - \frac{\partial \psi}{\partial p}; \ w = \frac{\partial \psi}{\partial y}
\]
Figure 13: A depiction of a kata cold front, with the dry intrusion (white arrows across cold front) overrunning the warm conveyor belt (large stippled arrow) and forming a split front. Taken from Browning and Monk (1982).

and $u$ is the along-frontal wind speed, $v$ is the across-frontal wind speed, $w$ is vertical velocity, the $g$ and $ag$ subscripts denote the geostrophic and ageostrophic components of the horizontal wind, $p$ is pressure, $\gamma = \frac{R}{f p_0} \left( \frac{p_0}{p} \right)^{c_v/c_p}$, $R$ is the gas constant for mixed air, $c_v$ and $c_p$ are the heat capacities of dry air at constant volume and pressure, $p_0$ is a reference value for pressure, $J_{yp}(u_1, u_2) = \frac{\partial u_1}{\partial y_p} \frac{\partial u_2}{\partial p} - \frac{\partial u_1}{\partial p} \frac{\partial u_2}{\partial y_p}$, $\dot{\theta} = \frac{d\theta}{dt}$, $F_x$ is the $x$-component of friction, and:

$$M_g = u_g - fy$$

is geostrophic momentum.

The lefthand side of the Sawyer–Eliassen equation quantifies the atmosphere’s response in terms of the ageostrophic streamfunction, $\psi$, to the frontogenetical forcing on the righthand side. The Sawyer–Eliassen equation illustrates that, for a given forcing, the resultant ageostrophic motion is dependent on the cross-frontal and vertical gradients of geostrophic momentum and potential temperature (the coefficients of the terms on the lefthand side). If these gradients are known, then a given forcing implies a partial differential equation for the ageostrophic streamfunction.
Bennetts and Hoskins (1979) suggested that frontal rainbands form when latent-heat release renders the atmosphere symmetrically unstable to saturated ascent. Symmetric stability collectively describes the coefficients on the lefthand side of the Sawyer–Eliassen equation.

The atmosphere may be gravitationally stable:

\[ \frac{\partial \theta}{\partial z} > 0 \]

and inertially stable:

\[ \frac{dM_g}{dy} > 0 \]

However, any parcel displacement (the righthand side of the Sawyer–Eliassen equation) results in a restoring force, attempting to return the parcel to its original \( \theta \) and \( M_g \) values (Houze 1993, p. 55). If \( \theta \) surfaces are steeper than \( M_g \) surfaces, as shown in Fig. 14, then a parcel displacement at an angle between those of the \( \theta \) and \( M_g \) surfaces means that the parcel finds itself in an environment of lower \( \theta \) and so continues to rise. Similarly, it finds itself in an environment of greater \( M_g \) and so is subject to a restoring force in the direction of lower \( M_g \) (to the left). Thus, the resultant restoring force pushes the parcel in the same direction as the displacement and it continues to rise in this slantwise fashion. This configuration between \( \theta \) and \( M_g \) surfaces is known as symmetric instability. If \( \theta \) is replaced by equivalent saturated potential temperature, \( \theta_{es} \), in the above argument then there is conditional symmetric stability (CSI) and describes the required state of the atmosphere for saturated parcels to ascend freely in a slantwise fashion. When a saturated parcel is lifted to its level of free slantwise convection (the level at which it can freely ascend slantwise), the CSI is released and slantwise convection occurs.

Convectively available potential energy (CAPE) diagnoses the potential for upright convection in a column of the atmosphere:

\[ \text{CAPE} = g \int_{\text{LFC}}^{\text{LNB}} \frac{\theta_{\text{par}}(z) - \theta_{\text{env}}(z)}{\theta_{\text{env}}(z)} dz \]

(Houze 1993, p. 283), where \( g \) is acceleration due to gravity, LFC and LNB are the levels of free convection and neutral buoyancy, respectively, and \( \theta_{\text{par}} \) and \( \theta_{\text{env}} \) are the potential temperature of the air parcel and environment, respectively. Analogously, slantwise convectively available potential energy (SCAPE) is calculated as the CAPE along an \( M_g \) surface and, thus, diagnoses the potential in the atmosphere.
for slantwise convection (Emanuel 1994).

The calculations of CAPE and SCAPE assume a steady-state background flow. For CAPE, this assumption is reasonable because the timescale of upright convection is far less than the timescale at which a baroclinic wave develops. However, due to the slower ascent associated with slantwise convection, the release of CSI takes place on a much longer timescale, more comparable with that of the background synoptic-scale flow. Therefore, the background flow may evolve significantly during the release of CSI. Gray and Thorpe (2001) showed that this problem may lead to substantial errors in the calculation of SCAPE and, thus, derived an extension of parcel theory to three dimensions, allowing for the evolution of the synoptic-scale flow during the release of CSI. This alternative diagnostic was found to produce significant differences in the distribution and magnitude of SCAPE within an extratropical cyclone used as a case study.

However, another difficulty in diagnosing the release of CSI is that upright and slantwise convection may occur simultaneously within a frontal zone (e.g., Browning et al. 2001), and interactions may occur between upright and slantwise convection, as follows. When slantwise convection occurs and stabilises the atmosphere to slantwise
ascent, gravitational instabilities may be generated further aloft (Xu 1986), a process known as downscale development. Also, the release of gravitational instability may be followed by the generation of CSI and hence slantwise ascent, a process known as upscale development. However, upright ascent followed by slantwise ascent does not imply the release of CSI. Slantwise ascent may occur from a process known as ΔM adjustment, which is when upright ascent results in inertial instability, whose release generates horizontal motion, tilting the updraft (e.g., Holt and Thorpe 1991; Pizzamei et al. 2005; Gray and Dacre 2008).

Another problem with diagnosing CSI arises from the definition of CSI as a region where moist geostrophic PV, $\frac{1}{\rho} \zeta_g \cdot \nabla \theta_{es}$, is negative, but the atmosphere is gravitationally and inertially stable (e.g., Schultz and Schumacher 1999). Moist geostrophic PV is often near zero within frontal rainbands (Emanuel 1988), implying near-neutral stability to slantwise convection. In such situations, whether CSI was present but has been released, or CSI was never present is uncertain. Even in model simulations, although evidence for CSI release may be exhibited, CSI cannot be conclusively diagnosed as a cause of precipitation bands (e.g., Knight and Hobbs 1988; Xu 1992; Zhang and Cho 1995).

Because of the many problems associated with diagnosing the release of CSI, the experiments conducted in this thesis are not analysed within a CSI framework. Nevertheless, the concept of negative moist PV is useful in that it illustrates how the atmosphere may be destabilised. The evolution of moist PV, following air parcels, is given by:

$$\frac{D q_w}{Dt} = \frac{f g}{\theta_0^2} \kappa \cdot (\nabla \theta_w \wedge \nabla \theta) + \frac{f g}{\theta_0} \zeta \cdot \nabla Q + \frac{f g}{\theta_0} F \cdot \nabla \theta_w$$

(Bennetts and Hoskins 1979), where $Q$ and $F$ are diabatic and frictional terms, and $\theta_w$ is wet-bulb potential temperature. However, $\theta_w$ should be replaced by $\theta_{es}$ in line with the above definition of moist PV. The first term describes changes in $q_w$ due to moisture gradients along trajectories. The second term is due to diabatic processes and the third due to frictional processes. Therefore, as air parcels moving along a conveyor belt within a cyclone encounter frictional and diabatic processes, moist PV anomalies are generated, affecting the stability (whether gravitational, inertial, symmetric, or some combination thereof). Therefore, these frictional and diabatic processes may have profound influences on the formation and maintenance of precipitation bands.
1.2.3 Frictional and diabatic effects on bands

Precipitation bands are profoundly affected by surface friction. Bands forced at the surface may be enhanced by frictional convergence (e.g., Bond and Fleagle 1985; Knight and Hobbs 1988; Bénard et al. 1992; Doyle 1997), which is greater for rougher surfaces, as is investigated throughout this thesis. Another effect of surface friction, as described in section 1.1.4, is the reduction of synoptic-scale forcing and rate of development of cyclones, which we may expect to affect bands indirectly. The sensitivity of bands to this synoptic-scale forcing was investigated by Gray and Dacre (2008), via idealised simulations of a cold front forced by deformation. They found that greater deformation strain causes an earlier onset of convection and better organisation of the convection. Greater low-level baroclinicity between simulations enhanced the $\Delta M$ adjustment for a given convective updraft, which may enhance multiple banding, as argued by Pizzamei et al. (2005). This thesis is not concerned with the sensitivity of bands to synoptic-scale forcing, but Papers 1 and 2 of this thesis indirectly address this relationship by increasing surface friction between idealised simulations (see sections 1.3.1 and 1.3.2), thereby reducing the synoptic-scale forcing near the surface.

The effects of diabatic processes on bands have been investigated in the literature in simple idealised models. For example, Thorpe and Emanuel (1985) found that latent-heat release in the updraft associated with a frontal rainband increases the rate of frontogenesis and narrows the updraft, intensifying the rainband. Knight and Hobbs (1988) found that, when a region of negative moist PV is prescribed above the surface cold front, multiple bands can form up the cold front. They noted that in the real atmosphere this region of negative moist PV could be created by diabatic processes.

Evaporative cooling in downdrafts may also affect bands. For example, Pizzamei et al. (2005) found that, for rainbands generated by slantwise ascent, evaporative cooling in the associated slantwise downdraft can generate a gravity current, which triggers the upright convection that generates a new rainband. This upright ascent is then subject to $\Delta M$ adjustment and hence can generate a further gravity current in the slantwise downdraft. Hence, this positive feedback can lead to multiple rainbands.

Surface sensible- and latent-heat fluxes may also affect bands by altering the lower-tropospheric lapse rate and moisture content (the latter potentially leading to enhanced latent-heat release within updrafts). Surface sensible heat flux can also
alter the temperature gradient near the surface, as described in section 1.1.4, which may also affect bands.

1.3 Aims and structure of this thesis

This PhD is funded by the Diabatic Influences on Mesoscale Structures in Extratropical Storms (DIAMET) project, whose goal is an improved understanding of how friction and diabatic processes (e.g., latent-heat release, and surface sensible- and latent-heat fluxes) affect high-impact mesoscale features, e.g., precipitation bands, within midlatitude weather systems. This thesis aims to generalise the observations of bands gained during the DIAMET field campaigns (Vaughan et al. 2014), and investigate the sensitivity of precipitation bands within baroclinic waves to frictional and diabatic processes. Therefore, this thesis will combine the studies investigating the sensitivity of the large-scale structure of idealised baroclinic waves, described in section 1.1.4, with the studies investigating the sensitivity of precipitation bands in simpler idealised models, described in section 1.2.3.

As argued in section 1.2.3, frictional and diabatic processes within baroclinic waves may play crucial roles in the formation and maintenance of bands, and, thus, may explain much of the variation from case to case. To address this hypothesis, the approach of the thesis is partly to examine precipitation-radar data (described in chapter 2.1) at times when bands formed, but predominantly to perform idealised and real-data simulations with the Weather Research and Forecasting (WRF) model (described in chapter 2.2) to investigate the formation and sensitivity of bands.

The use of WRF, a primitive-equation model, as opposed to a simpler model, allows us to simulate realistic precipitation bands, as can be gauged by a comparison of simulated bands to those observed by precipitation radar (Fig. 15). Unlike with observational data, with a model we can examine the physics leading to the formation of various types of bands and, thus, gain a greater understanding of how bands are generated in the real atmosphere. Furthermore, by performing multiple simulations, in which the lower boundary of the model domain is altered and the physics are edited between simulations, we can observe and quantify the effects that the various diabatic influences (including friction) have on bands. Thus, we can gain a greater understanding of why bands form in some situations, but not in others, and why bands are qualitatively and quantitatively different between cases.

This thesis is presented in alternative format, meaning that the results of the thesis are presented in the form of three journal articles of which I am the lead
Figure 15: A comparison of a snowband at 18 UTC 30 Nov 2010, in terms of precipitation rate (mm h\(^{-1}\)) at 1-km grid spacing: (a) Met-Office precipitation-radar composite; (b) Simulation with WRF model. Adapted from Fig. 15 of Paper 3 of this thesis.
As the lead author, I performed all the simulations in the papers, collected and processed all the additional data presented (with the exception of the synoptic composites presented in Fig. 11 of Paper 3, which was done by my co-author David M. Schultz), led the analysis of the results, wrote the drafts, and, in the case of Papers 1 and 3, co-ordinated the responses to reviewers. The contributions of my co-authors were to aid the analysis of the results, suggest future directions for the papers, and provide comments on drafts, which improved subsequent drafts.

The alternative-thesis format allows the work that has been done for this PhD to be divided into three bodies of work, which are distinct, with each focussing on a specific problem within this subject matter, but which collectively highlight the sensitivity of precipitation bands to diabatic processes. In this thesis, the first journal article investigates the sensitivity to diabatic processes of different band types together in baroclinic waves. The second and third journal articles then focus on two specific band types, motivated by recent observations, which are associated with high-impact weather.

### 1.3.1 Paper 1

This first paper within this alternative thesis is designed to start the thesis as generally as possible, within the subject material. The idea is to make no assumptions of what bands may form within baroclinic waves and to investigate band formation, simply by applying the baroclinic-wave theory detailed in section 1.1. Therefore, the WRF model is used with a moist idealised baroclinically unstable midlatitude jet as the initial condition (section 1.1.4). The advantage of the idealised-modelling approach is that the complications in the real atmosphere that may affect bands (e.g., non-uniform initial conditions, case-to-case variability, and influences of weather systems in other parts of the world) can be removed and the natural tendency of a baroclinic wave to produce various types of bands can be isolated.

This paper also removes the complications in the lower boundary (e.g., orography, coastlines, and thermal and frictional nonuniformities) that may affect bands in the real atmosphere. Therefore, an all-ocean lower boundary is used (hence flat and without coastlines) with constant roughness length (the theoretical height above the surface at which wind speed is zero, depending on surface drag) and a zonally uniform SST distribution. By performing multiple simulations with varying roughness length, latent-heat release switched on and off, and the SST varied between simulations, this paper explores the diabatic factors affecting precipitation-band formation.
in baroclinic waves. These simulations are performed at 20-km grid spacing to
resolve bands spaced hundreds of km apart.

This paper is entitled “Precipitation banding in idealized baroclinic waves” and
presented in this thesis in the form in which it was resubmitted to Monthly Weather
Review as part of the DIAMET collection (http://journals.ametsoc.org/page/diamet),
after making amendments requested by the reviewers.

1.3.2 Paper 2

Having given an overview of all the bands that can be produced by an idealised
baroclinic wave at a relatively course resolution in Paper 1, Papers 2 and 3 focus on
two specific band types (Fig. 1) and employ higher-resolution simulations. Paper
2 is a natural progression from Paper 1, in that it employs further WRF idealised
baroclinic-wave simulations to simulate bands that are not resolved in simulations at
20-km grid spacing, specifically precipitation cores (PCs), where the heaviest rainfall
along and ahead of the cold front occurs (Fig. 1a).

Previous literature has frequently proposed that PCs form from horizontal-shear
instability (see introduction to Paper 2), and Kawashima (2011) simulated precipita-
tion structures, resembling observed PCs, in an idealised shear–deformation model,
showcasing the dependence of these structures to horizontal shear, vertical shear,
and static stability. The approach of Paper 2 is to show, in a more realistic model,
the sensitivity of PCs to diabatic factors, and whether the differences in PCs be-
tween simulations are due to the variations in shear and static stability, as found by
Kawashima. As in Paper 1, the diabatic factors considered are roughness length,
latent-heat release, and SST.

These PCs form in the real atmosphere along mature cold fronts. Therefore,
one of the simulations in Paper 1, after the cold front has formed, is re-initialised
to produce the initial condition for all simulations in this study. Although the
simulations in Paper 1 are performed at 20-km grid spacing to resolve bands spaced
on the order of hundreds of km apart, in Paper 2, a nested domain of 4-km grid
spacing is inserted to resolve the PCs spaced tens of km apart, which is shown to
be sufficient. This is a limited study to investigate the precipitation structures that
appear when the model is run at higher resolution than in Paper 1, and to apply
the methodology of Paper 1 to understanding the physics behind these structures.

Paper 2 is entitled “Precipitation cores along a narrow cold-frontal rainband in
idealized baroclinic waves”. As of the date of submission of this thesis, this paper
has not yet been submitted to a journal, but, with Paper 1, it will be submitted to *Monthly Weather Review* as part of the DIAMET collection.

### 1.3.3 Paper 3

Finally, Paper 3 focuses on postfrontal bands. In the winters of 2009–10 and 2010–11, there were several weeks in which the UK was affected by severe cold-air outbreaks from the north and east (Fig. 7b). The impacts of these cold-air outbreaks are described in the introduction to Paper 3. During the cold-air outbreaks, precipitation-radar data revealed that the extreme snowfall over the British Isles regularly fell in the form of bands, most notably over the English Channel and Irish Sea and were aligned along the major axis of each body of water (e.g., Fig. 1b).

These two winters presented a fairly small and manageable dataset with which to investigate extreme postfrontal bands that can occur when extremely cold air aloft interacts with the lower boundary. Because this paper focuses on a more specific set of bands than the other papers in this thesis, there is a greater observational component to this paper. This observational component consists of a climatology of precipitation-radar and sounding data, thus revealing the state of the atmosphere at the times that bands were and were not observed.

As in Papers 1 and 2, the WRF model is also used to investigate the physics associated with band formation and sensitivity to diabatic factors. In this paper, real-data, rather than idealised, simulations are employed because, in this paper, the aim is to determine why these particular bands formed. Also, as shown in the paper, these bands were reliant on the complications in the lower boundary that Papers 1 and 2 neglect, so real-data simulations allow an investigation of how the particular topography of the British Isles and northern France (orography and coastlines) interacted with the cold air aloft to produce the snowbands. This thesis is not concerned with the influence of orography on precipitation bands. However, in this paper, orography is one of the influences that is investigated, via the sensitivity simulations, just to check that orography was not a dominant factor in the snowbands and that the bands were indeed more dependent on the diabatic processes of interest to this thesis.

Thus, Paper 3 provides a climatology and model-sensitivity experiments of these UK bands, which has not been done previously in the literature. This paper has been published in the *Quarterly Journal of the Royal Meteorological Society*, entitled “Snowbands over the English Channel and Irish Sea during cold-air outbreaks”. The
paper is presented in this thesis in its published form. However, Figs. 11, 13, 14, and 15 of the paper, which were full page, have been shrunk from the typesetting, which may make these figures difficult to view when printed, so the original full-page versions of these figures are appended at the end of the thesis (Figs. 21, 22, 23, and 24).
2 Methods

This thesis investigates precipitation bands via observations and model simulations. The bands are observed via Met Office precipitation-radar data (Met Office 2009, described in chapter 2.1) and simulated with version 3 of the WRF model (Skamarock et al. 2008, described in chapter 2.2).

2.1 Met Office precipitation-radar data

The Met Office’s precipitation-radar data is generated by a network of eighteen radars, evenly distributed around the British Isles (Fig. 16). Each radar emits an electromagnetic pulse at 5.6 cm wavelength, which is reflected by precipitation droplets (Fig. 17), including rain, snow, and hail. The intensity of the signal returned, the reflectivity, indicates the density of droplets and, thus, the intensity of precipitation. The time taken for a pulse to return indicates the distance of precipitation from the radar. The data from each radar are sent to the Met Office, where an algorithm is performed to combine the data from each radar and eliminate various errors (see below). A single image is thus produced of the reflectivity ($Z$, in $\text{mm}^6\text{m}^{-3}$) at 1-km grid spacing throughout the British Isles at a given moment in time, which is converted into estimated precipitation rate ($R$, in $\text{mm} \text{h}^{-1}$), using the equation in Table 1 of Harrison et al. 2009:

$$Z = 200R^{1.6}$$  \hspace{1cm} (7)

Thus, precipitation radar estimates the instantaneous distribution and intensity of precipitation reaching the surface. The morphology of precipitation features, specifically bands, can therefore be visualised and compared with simulated bands (e.g., Fig. 15).

The precipitation-radar data are just an estimate of the distribution and intensity of precipitation and there are a number of errors that may occur during the production of any given image.

- Precipitation radars emit pulses at non-zero elevation angles (typically between $0.5^\circ$ and $4^\circ$) in order to avoid ground clutter, e.g., hills and trees. Combined with the curvature of the earth, this means that, with increasing distance from the radar, the pulse is reflected increasingly high above the ground. Between the level of reflection and the ground, the precipitation intensity may increase.
Figure 16: The locations of the radars that contribute to the Met Office precipitation-radar data, indicated by the red dots. The inner, middle, and outer circles around each radar indicate coverage at 1-km, 2-km, and 5-km grid spacing, respectively. Taken from Met Office (2009).
Figure 17: A depiction of a precipitation radar emitting an electromagnetic pulse, which is reflected by precipitation droplets and returned to the radar. Taken from Met Office (2009).

(e.g., because the pulse is reflected at the cloud level or above, or because precipitation is enhanced by orography nearer the ground) or decrease (e.g., due to evaporation). Therefore, at long ranges, the estimation of precipitation rate may differ significantly to the precipitation rate at the surface.

- Different precipitation types are calibrated differently to allow for the difference in density between droplets. However, when a pulse is reflected at the level of the atmosphere at which snowflakes are melting, the bright reflective surface of the snowflakes sends back a spurious high echo, overestimating precipitation intensity. This error is known as the bright band.

- Spurious echoes may be produced by aircraft contrails, insects, interference from other radars, and other features in the atmosphere, giving false estimates of precipitation. However, these echoes are normally short lived and can be easily differentiated from genuine precipitation echoes.

- Different clouds have different sizes of droplets. For example, convective clouds have larger droplets than stratiform clouds. Because radar echoes do not indicate the size of droplets, a different average droplet size is assumed for each cloud type (e.g., greater for convective clouds than for stratiform clouds) in order to estimate precipitation rate from radar echoes. This approach leads
to the radar underestimating precipitation rate when detecting smaller-than-average droplets and overestimating precipitation rate when detecting larger-than-average droplets. However, when the echoes are averaged over a grid box of the radar data, these errors are reduced.

- Pulses are refracted when passing through air of varying density. This effect occurs most commonly when the pulse passes through an inversion layer above the boundary layer, in which case the pulse is refracted downwards and reflected off the ground, so that a large spurious echo is returned to the radar. However, this effect mostly occurs in stable high-pressure conditions, in which case isolated areas of extreme precipitation are unlikely and so the echoes are easily identified as spurious.

- Although the pulses are emitted at non-zero elevation angles, some tall structures, such as buildings and mountains, reflect pulses, leading to spurious echoes. However, on cloudless days, the echoes returning from these structures can be identified and then removed from the radar data on these and other days.

Thus, except where radar echoes increase or decrease between the level of reflection and the ground, and the bright-band echoes, the errors arising from the above are normally detected and corrected. Paper 3, the only paper in this thesis to use precipitation-radar data other than for motivation, acknowledges these two potential sources of error, but explains why they are unlikely to affect the conclusions of the paper (section 2.1 of Paper 3).

2.2 The Weather Research and Forecasting model

WRF is an Eulerian, compressible, non-hydrostatic primitive-equation model. The WRF output shown in this thesis is a product of the co-ordinate system (section 2.2.1), the governing equations (section 2.2.2) and their discretisation (section 2.2.3), the initial (section 2.2.4) and boundary (sections 2.2.5 and 2.2.6) conditions, and the parameterised sub-gridscale processes (section 2.2.7).

2.2.1 Co-ordinates

The vertical co-ordinate is a terrain-following hydrostatic-pressure co-ordinate, \( \eta = (p_h - p_{ht})/\mu \), where \( \mu = p_{hs} - p_{ht} \); \( p_h \) is the hydrostatic pressure of the dry atmosphere,
Figure 18: The terrain-following hydrostatic-pressure coordinate, $\eta$, that is used as WRF’s vertical coordinate. Taken from Skamarock et al. 2008.

and $p_{hs}$ and $p_{ht}$ are $p_h$ at the surface and model top, respectively (Fig. 18). Therefore $\eta$ ranges from 1 at the surface to 0 at the model top.

Papers 1 and 2 of this thesis are of idealised baroclinic-wave simulations, in which the horizontal co-ordinates are $(x,y)$, where the initial jet is constant in the $x$-direction. In Paper 3, real-data simulations are performed, in which a Lambert conformal mapping is used to map the latitude–longitude grid to an $x$-$y$ grid (Fig. 12 of Paper 3).

2.2.2 Governing equations

The time stepping in all the WRF simulations in this thesis is performed by the Advanced Research WRF (ARW) dynamical core. The ARW solves the flux-form
variables:

\[ \mathbf{V} = (U, V, W) = (\mu u, \mu v, \mu w) \quad \Omega = \mu \phi \quad \Theta = \mu \theta \]

together with \( \phi = gz \) (geopotential), \( p \) (pressure), and \( \rho \) (density). In addition, there is an equation for \( Q_1, Q_2, \ldots \), where \( Q_i = \mu q_i \) and \( q_1, q_2, \ldots \) are the mixing ratios (kg of water per kg of dry air) of each microphysical species (water vapour, cloud water, ice, rain water, etc.) that the chosen microphysics scheme (section 2.2.7) includes. Thus, the moist flux-form Euler equations are:

\[ \frac{\partial U}{\partial t} + (\nabla \cdot \mathbf{V} u) + \rho \frac{1}{\rho} \frac{\partial p}{\partial x} + \rho_d \frac{\partial p}{\partial \eta} \frac{\partial \phi}{\partial x} = F_U \quad (8) \]
\[ \frac{\partial V}{\partial t} + (\nabla \cdot \mathbf{V} v) + \rho \frac{1}{\rho} \frac{\partial p}{\partial y} + \rho_d \frac{\partial p}{\partial \eta} \frac{\partial \phi}{\partial y} = F_U \quad (9) \]
\[ \frac{\partial W}{\partial t} + (\nabla \cdot \mathbf{V} w) - g \left( \rho_d \frac{\partial p}{\partial \eta} - \mu \right) = F_W \quad (10) \]
\[ \frac{\partial \Theta}{\partial t} + (\nabla \cdot \mathbf{V} \theta) = F_\theta \quad (11) \]
\[ \frac{\partial \mu}{\partial t} + (\nabla \cdot \mathbf{V}) = 0 \quad (12) \]
\[ \frac{\partial \phi}{\partial t} + \frac{1}{\mu} [(\mathbf{V} \cdot \nabla \phi) - g W] = 0 \quad (13) \]

and, for each \( q_i \):

\[ \frac{\partial Q_i}{\partial t} + (\nabla \cdot \mathbf{V} q_i) = F_{Qi} \quad (14) \]

together with the diagnostic relation for dry density:

\[ \frac{\partial \phi}{\partial \eta} = -\frac{\mu}{\rho_d} \quad (15) \]

and the equation of state for full pressure (vapour plus dry air):

\[ p = p_0 \left( \frac{\rho_d R_d \theta_m}{p_0} \right)^{c_p/c_v} \quad (16) \]

where \( \rho_d \) is the density of dry air, \( \theta_m = \theta \left( 1 + R_v q_v \right) \), \( q_v \) is the mixing ratio of water vapour, \( g \) is acceleration due to gravity, \( c_p \) and \( c_v \) are the heat capacities of dry air at constant pressure and volume, \( R_d \) and \( R_v \) are the gas constants for dry air.
and water vapour, and \( p_0 \) is a reference pressure. In addition, for some prognostic variable \( \Phi \):

\[
\nabla \cdot \mathbf{V} \Phi = \frac{\partial (U \Phi)}{\partial x} + \frac{\partial (V \Phi)}{\partial y} + \frac{\partial (\Omega \Phi)}{\partial \eta}
\]

\[
\mathbf{V} \cdot \nabla \Phi = U \frac{\partial \Phi}{\partial x} + V \frac{\partial \Phi}{\partial y} + \Omega \frac{\partial \Phi}{\partial \eta}
\]

and \( F \) is the forcing for \( \Phi \) from model physics, turbulent mixing, spherical projections, and the earth’s rotation.

### 2.2.3 Discretisation

These equations are discretised in time, using the Runge–Kutta 3 time-integration scheme (Wicker and Skamarock 2002). This scheme consists of re-arranging the prognostic model equations (8, 9, 10, 11, 12, 13, 14) as \( \frac{\partial \Phi}{\partial t} = R(\Phi) \), where \( R(\Phi) \) consists of all the terms other than the time derivative. The solution of a prognostic variable at time \( t \), \( \Phi(t) \), is then advanced to \( \Phi(t + \Delta T) \), where \( \Delta T \) is the time step, by the following three steps:

\[
\Phi^* = \Phi(t) = \frac{\Delta t}{3} R(\Phi(t))
\]

\[
\Phi^{**} = \Phi(t) = \frac{\Delta t}{2} R(\Phi^*)
\]

\[
\Phi(t + \Delta t) = \Phi(t) = \Delta t R(\Phi^{**})
\]

The spatial derivatives in the governing equations are discretised by an Arakawa C-grid staggering, in which wind variables are defined at the edges of grid boxes and all other variables are defined in the centre of grid boxes (Fig. 19). Any given spatial derivative, \( \frac{\partial \Phi}{\partial x} \), is numerically represented by:

\[
\delta_x \Phi = \frac{\Phi_{i+\frac{1}{2}} - \Phi_{i-\frac{1}{2}}}{\Delta x}
\]

(or equivalently, replacing \( x \) by \( y \) or \( \eta \), and \( i \) by \( j \) or \( k \)), where \( \Delta x \) is the horizontal grid spacing and, for horizontal derivatives:

\[
\Phi_{i+\frac{1}{2}} = \frac{\Phi_i + \Phi_{i+1}}{2}
\]

because \( \Delta x \) and \( \Delta y \) are constant (Fig. 19a). However, this representation is not appropriate for vertical derivatives because \( \Delta \eta \) is not constant (Fig. 19b). Therefore,
Figure 19: The Arakawa C-grid with wind variables defined at the edge of grid boxes ($u, v,$ and $w$ on the zonal, meridional, and vertical edges, respectively) and all other variables defined in the centre of grid boxes. $\Delta x$, $\Delta y$, and $\Delta \eta$ are the grid spacings in the $x$, $y$, and $\eta$ directions, where $\Delta x$ and $\Delta y$ are constant, and $\Delta \eta$ varies between each vertical level. Taken from Skamarock et al. (2008).

the operator $\Phi_{k+\frac{1}{2}}$ vertically interpolates variables on mass levels, $k$ (the centres of grid boxes), to $w$ levels, $k + \frac{1}{2}$ (the edges of grid boxes):

$$\Phi_{k+\frac{1}{2}} = \frac{1}{2} \left( \frac{\Delta \eta_k}{\Delta \eta_{k+\frac{1}{2}}} \Phi_{k+1} + \frac{\Delta \eta_{k+1}}{\Delta \eta_{k+\frac{1}{2}}} \Phi_k \right)$$  

(22)

where $\Delta \eta_{k+\frac{1}{2}} = \frac{\Delta \eta_k + \Delta \eta_{k+1}}{2}$

2.2.4 Initial conditions

The initial condition for any WRF simulation is generated by the WRFinput routine. The initial condition for the idealised baroclinic-wave simulations in Papers 1 and 2 is a moist midlatitude jet in thermal wind balance, spanning 8000 km from south to north (Fig. 20). The jet is produced by prescribing a baroclinically unstable PV distribution in the $y$-$z$ plane (as in Fig. 2, before the perturbations are made) and, under the assumption of thermal wind balance, inverting the PV distribution (section 1.1) to obtain the distributions of $\theta$ and $u$, and produce a two-dimensional input-jet binary file. WRFinput then uses this input jet to generate
Figure 20: The WRF initial jet. (a) Surface $\theta$ (coloured every 5 K) and 300-hPa wind vectors (m s$^{-1}$); (b) Cross section from south to north, as indicated by bold line in panel a, showing $u$ (labelled colours, m s$^{-1}$), $\theta$ (red contours every 5 K), and water-vapour mixing ratio (labelled greys, g of water vapour per kg of dry air). Domain extends 1000 km further equatorward and 2000 km further poleward than area shown.

The three-dimensional initial condition, depending on the user’s choice of horizontal grid spacing, the west–east span of the domain, the number of vertical levels, and the height of the model top. WRFinput also generates a distribution of water-vapour mixing ratio, decreasing from equator to pole and from the surface to the the model top (Fig. 20b), based on the pressure levels and $\theta$ distribution (with all other microphysical species initially zero everywhere).

The initial condition for the real-data simulations in Paper 3 is generated from an analysis of the European Centre for Medium Range Weather Forecasts (ECMWF) operational model (ECMWF 1994) at 0.5° latitude–longitude grid spacing. The WRF Preprocessing System (WPS) takes this data and generates appropriately formatted initialisation data for WRF. WPS consists of the following routines:
• **Geogrid** takes time-invariant two-dimensional geographical data over the latitude and longitude range that the user has chosen for the simulation (e.g., terrain height, whether land or sea, vegetation type, roughness of the surface, albedo) and interpolates this data onto a domain defined by the latitude-longitude range, the type of mapping projection, and the horizontal grid spacing. Geogrid also generates any nested domains, defined by the area that each nested domain occupies within its parent domain and the nested domain’s grid spacing.

• **Ungrib** takes the Gridded Binary (GRIB) file containing the model analyses that are being used to initialise WRF and provide lateral boundary conditions (section 2.2.6) and extracts the three-dimensional meteorological variables that the ARW solves (section 2.2.2). These analyses are available every six hours and are required for the duration of the simulation. Ungrib thus produces a file of all the required variables for each six-hour period throughout the WRF simulation. In the simulations in Paper 3, National Oceanographic and Atmospheric Administration (NOAA) SST-analysis data (Reynolds 1988) at 0.5° grid spacing are also used. Thus, Ungrib extracts SST from this data and produces a SST file every six hours.

• **Metgrid** interpolates the meteorological variables obtained by Ungrib onto each of the domains generated by Geogrid, producing a Metgrid file for each domain every six hours.

WRFinput then maps the data in the Metgrid files to the vertical levels over which the WRF simulation is to be run, depending on the user’s choice of the number of vertical levels and the model-top pressure, to generate the initial and lateral boundary conditions.

### 2.2.5 Lower- and upper-boundary conditions

In all WRF simulations, the lower boundary is the earth’s surface, at which the free-slip boundary condition is applied. In addition, momentum, heat, and moisture fluxes from the surface are provided by the surface-layer parameterisation (section 2.2.7). The upper boundary condition is constant pressure, $p_{ht}$, at the model top (Fig. 18), which is specified by the user for real-data simulations (50 hPa in Paper 3) and, in the idealised baroclinic-wave simulations, calculated by WRFinput, based on the specified model-top height of the domain and the number of vertical
levels. In Papers 1 and 2, the model top is 16 km and there are 80 vertical levels, which leads to $p_{ht} = 108$ hPa.

### 2.2.6 Lateral boundary conditions

For real-data simulations, the lateral boundary conditions are, for the outer domain, provided by the model-analysis data (section 2.2.4), which extend beyond the latitude–longitude range of the WRF simulation. Because these analyses are only available every six hours, WRFinput interpolates these boundary conditions to provide boundary conditions every time step.

For the idealised baroclinic-wave simulations, a periodic boundary condition is applied in the $x$-direction, so that, $\Phi(x + nL, y, z) = \Phi(x, y, z)$ for some variable $\Phi$, where $n$ is any integer and $L$ is the width in the $x$-direction, or wavelength, of the domain (4000 km in Papers 1 and 2). A symmetric boundary condition is applied in the $y$-direction, so that $v = 0$ at the boundaries and, either side of the boundary, $v(x, y_b - y, z) = -v(x, y_b + y, z)$ and, for all other variables, $\Phi(x, y_b - y, z) = \Phi(x, y_b + y, z)$.

For both real-data and idealised simulations, the lateral boundary conditions for any nested domain (as in Papers 2 and 3) are provided by the output of the nest’s parent domain. Because the time step of the parent domain is greater than for the nested domain, these boundary conditions are interpolated to provide boundary conditions every time step for the nest.

### 2.2.7 Parameterisations

The ARW parameterises various sub-gridscale physical processes, which contribute to the righthand side of equations 8, 9, 10, 11, and 14. These contributions are known as tendencies. The parameterisations thus change the momentum, heat, and moisture content of air parcels. The parameterisations employed in this thesis are as follows:

- **Microphysics** parameterises the change of state of water between different microphysical species (the $q_i$ in equation 14), for example, the change from water vapour to cloud water, resulting in cloud formation, and the change from cloud water to rain water, resulting in rainfall. The latent heating or cooling associated with each phase change is also calculated. WRF employs bulk microphysics schemes, meaning that a function is used, such as exponential or gamma functions, relating the size to the density of droplets, for a given
species, which can be normalised and integrated over a complete size distribution (Straka 2009, p. 3). The scheme thus predicts the mixing ratio of each microphysical species, at each time step. Some schemes, known as double-moment schemes, as used in this thesis, also predict the number concentration (number of droplets per kg) of certain species, in which case an extra equation is solved every time step by the ARW for the number concentration of each species.

- **Cumulus** parameterisation calculates vertical fluxes of temperature and moisture, representing unresolved updrafts and downdrafts. The cumulus scheme is activated when the atmosphere becomes absolutely unstable in a column of the atmosphere, or conditionally unstable if saturated, and calculates the resulting profiles of temperature and moisture in that column. Entrainment into and detrainment out of the column are also calculated.

- **Radiation** is parameterised to represent incoming and outgoing radiation in a column, with separate schemes for longwave (e.g., infra-red) and shortwave (e.g., visible) radiation. Incoming radiation is calculated by the time of year, time of day, and latitude. To calculate the amount reaching the surface, absorption, reflection, and scattering are represented, depending on cloud cover and other microphysical species. The effects on these processes of other atmospheric constituents, such as O₃ and CO₂, are also calculated, using pre-defined tables of each constituent’s distribution. Longwave radiation reflected by the surface depends on land-use type (prescribed by Geogrid) and skin temperature. Reflected shortwave radiation depends on albedo (prescribed by Geogrid). Outgoing radiation is then also subject to absorption, reflection, and scattering.

- **Surface-layer** parameterisation calculates fluxes of momentum from the lower boundary into the boundary layer, depending on the roughness of the surface, and the magnitude and orientation of the surface winds. At sea grid points, the surface-layer scheme also calculates heat and moisture fluxes, depending on the SST and the magnitude and orientation of the surface winds.

- **Land-surface** parameterisation takes information from the microphysics, convection, radiation, and surface-layer schemes to calculate heat and moisture fluxes at land grid points. The fluxes are calculated by exchanging heat and moisture between multiple soil layers, and accounting for such processes as
evapotranspiration, soil drainage and runoff, depending on the vegetation type (prescribed by Geogrid).

- **Boundary-layer** parameterisation produces the fluxes of momentum, heat, and moisture from the surface-layer and land-surface schemes into the free atmosphere, which occur by eddy transport in the real atmosphere. These fluxes occur in a column the full depth of the atmosphere, not just in the boundary layer. The boundary-layer scheme calculates the flux profile in the boundary layer and stable layer, thereby providing tendencies throughout the column.

All of the above are employed in the real-data simulations in Paper 3 (although some aspects of the surface-layer parameterisation are removed in certain simulations, see section 6.1 of Paper 3). In the idealised simulations in Papers 1 and 2, there is no radiation or land-surface parameterisation (so no diurnal effects) and different simulations employ different combinations of the other parameterisations (see section 2 of Papers 1 and 2). For all simulated domains in this thesis at 5-km or less grid spacing, cumulus parameterisation is switched off because updrafts and downdrafts should be resolved at this grid spacing. For all domains with greater grid spacing, cumulus parameterisation is switched on, except when latent-heat release is subtracted from the microphysics\(^1\), for which the model requires cumulus parameterisation to be switched off.

Different parameterisation schemes are available for each of the above, of varying levels of sophistication (e.g., the number of soil layers in a land-surface scheme and the number of microphysical species represented in a microphysics scheme). In general, this thesis employs the more sophisticated schemes, and, where applicable, the schemes most appropriate for these particular simulations, due to the geographical location, grid spacing, etc. The particular schemes employed in each paper are summarised in Table 1 and work as follows:

- **Thompson microphysics** is a bulk microphysics scheme, which explicitly predicts the mixing ratios of water vapour, cloud water, rain water, cloud ice, graupel, and snow. The scheme is double-moment with respect to cloud ice, so that the number concentration, as well as mixing ratio, is predicted for cloud ice, but single-moment with respect to all other microphysical species. There-

\(^1\)Wherever in this thesis it is stated that latent-heat release is subtracted from the microphysics, the latent heat of all phase changes (not just condensation and evaporation) is set to zero.
fore, the scheme more accurately predicts ice processes and hence snowfall than fully single-moment schemes.

- **The Kain–Fritch cumulus scheme** checks, at each vertical level, whether there is grid-resolved vertical motion that may lead to moist convection. An air parcel is assigned a temperature perturbation, obtained by subtracting a threshold vertical velocity from the grid-resolved vertical velocity. If the resulting air-parcel temperature is greater than the environmental temperature at the parcel’s lifting condensation level (LCL), then the parcel is released at its LCL with its unperturbed temperature at a vertical velocity based on the perturbation temperature. The parcel’s convective vertical velocity is then estimated at each vertical level, using the same method, but including the effects of entrainment into, detrainment out of, and water loading of the parameterised updraft.

- **The Dudhia scheme** calculates a downward integration of shortwave radiation in a single spectral band, where the amount received at the surface is subject to clear air scattering, water vapour absorption, and cloud albedo and absorption.

- **The RRTM scheme** represents longwave radiation processes using 16 spectral bands, depending on cloud cover, CO$_2$ and O$_3$.

- **The MM5 Monin–Obukhov surface-layer scheme** calculates the surface exchange coefficients of momentum, heat, and moisture fluxes, depending on the frictional and thermal properties of the lower boundary, and the wind and temperature at the lowest model level. Heat and moisture fluxes are enhanced by a convective velocity, depending on the vertical thermal gradient near the surface.

- **The Noah land-surface scheme** uses four soil layers, extending to two metres below the surface, at each of which temperature, water, and ice are predicted. Depending on these parameters and the vegetation properties at the relevant grid box, root zone, evapotranspiration, soil drainage, and runoff are also calculated. The heat and moisture fluxes between the atmosphere and the land surface are thus calculated at each time step.

- **The Yonsei University boundary-layer scheme** identifies the boundary-layer height as the level of minimum flux in the inversion layer. The fluxes
throughout a column at the subsequent time step are then calculated, depending on the gradients of temperature, momentum, and moisture between this boundary-layer top and the surface, taking into account the surface fluxes calculated by the surface-layer and land-surface schemes.

Table 1: Parameterizations employed in each paper of this thesis

<table>
<thead>
<tr>
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<th>Paper 1</th>
<th>Paper 2</th>
<th>Paper 3</th>
</tr>
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<tbody>
<tr>
<td>Microphysics</td>
<td>Thompson (Thompson et al. 2008)</td>
<td>Kain–Fritsch (Kain 2004)</td>
<td></td>
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<tr>
<td>Cumulus</td>
<td></td>
<td></td>
<td></td>
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<tr>
<td>Shortwave radiation</td>
<td>-</td>
<td>-</td>
<td>Dudhia (Dudhia 1989)</td>
</tr>
<tr>
<td>Longwave radiation</td>
<td>-</td>
<td>-</td>
<td>Rapid Radiative Transfer Model (RRTM, Mlawer et al. 1997)</td>
</tr>
<tr>
<td>Surface layer</td>
<td>MM5 Monin–Obukhov (Monin and Obukhov 1954)</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Land surface</td>
<td>-</td>
<td>-</td>
<td>Noah (Niu et al. 2011)</td>
</tr>
<tr>
<td>Boundary layer</td>
<td>Yonsei University (Hong et al. 2006)</td>
<td></td>
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</tbody>
</table>

The above are used in at least one simulation and domain in the given papers. See section 2 of Papers 1 and 2, and section 6.1 of Paper 3, for simulations in which some parameterisations are switched off, and some aspects of microphysics and surface-layer parameterisation are subtracted.
3 Paper 1: Precipitation banding in idealized baroclinic waves.
Precipitation banding in idealized baroclinic waves

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ABSTRACT

Idealized baroclinic-wave simulations are presented to show the development of precipitation bands from a zonally uniform initial midlatitude jet above an ocean surface. For a frictionless lower boundary and with no latent-heat release or surface fluxes, warm advection is strong and a bent-back warm front forms. Although a narrow vertical-velocity maximum forms within the area of synoptic-scale ascent, near the triple point, only a wide warm-frontal band forms. As friction increases between simulations to that of an ocean, then a land surface, warm advection is reduced and the cold front becomes stronger, relative to the warm front. A separate narrow rainband forms along the cold front, which is more intense and farther removed from the wide warm-frontal band when friction is greater. In the simulation with ocean friction, after the narrow band decays, the precipitation becomes oriented along the warm conveyor belt in the warm sector. When latent-heat release is included, this warm-sector precipitation evolves into multiple bands, which eventually weaken with the cyclone. When surface heat and moisture fluxes are included, the ascent at the surface cold front stays strong and a well-defined cold front of the anafront variety persists through this mature stage. This surface-based ascent underlies the mid-level ascent associated with the multiple warm-sector bands and the surface precipitation remains in a single intense band. Therefore, strong fluxes inhibit multiple bands, but a simulation with lower sea-surface temperature (SST) more closely resembles the simulation without surface fluxes, demonstrating that the detailed structure and evolution of precipitation banding is sensitive to SST.

1. Introduction

Heavy precipitation within extratropical cyclones is often organized into mesoscale bands (e.g., Houze et al. 1976; Hobbs 1978; Houze and Hobbs 1982; Browning 1986, 1990, 2005; Browning et al. 1997; Browning and Roberts 1999). Houze and Hobbs (1982) classified precipitation bands by where they occur in extratropical cyclones, producing a conceptual model featuring six different types of precipitation bands: warm-frontal, warm-sector, wide cold-frontal, narrow cold-frontal, prefrontal cold-surge, and postfrontal bands. Browning (1986) showed that many of these types of precipitation bands were associated with airstreams within extratropical cyclones called conveyor belts (e.g., Carlson 1980). Thus, the Houze and Hobbs (1982) conceptual model serves as a template against which both observational and modeling studies can categorize bands within extratropical cyclones.

However, not all of these six types of bands appear within all cyclones. For example, Parsons and Hobbs (1983) documented five cyclones over the eastern North Pacific Ocean, all of which exhibited wide and narrow cold-frontal rainbands, but none of which exhibited warm-frontal bands. Conversely, a climatology of bands in cyclones over the northeast United States by Novak et al. (2004) found a predominance of bands east and northwest of the cyclone center, aligned along the warm front in the lower- and mid-troposphere. In addition, precipitation sometimes remains within a single wide band along a front, albeit with multiple lines of embedded features (Fig. 1a). But often, even if only a single front is analyzed, the frontal precipitation separates into multiple parallel bands of similar width and intensity (Fig. 1b). These examples demonstrate the considerable variety of precipitation structures within extratropical cyclones.

Previous research has identified a number of factors controlling the structure and evolution of banded features along fronts. For example, nonuniformities in the initial temperature gradient may result in the production of multiple fronts in an idealized baroclinic wave (Hoskins et al. 1984). However, such nonuniformities are not required to produce multiple bands. Even in two-dimensional (e.g., Knight and Hobbs 1988; Bénard et al. 1992a; Xu 1992;...
Pizzamei et al. 2005) and three-dimensional (e.g., Zhang and Cho 1995; Gray and Dacre 2008) idealized-model simulations, fronts with an initial single maximum of vertical velocity may develop multiple maxima of vertical velocity over time.

These observations and modeling results motivate the questions of what causes precipitation bands in cyclones and what determines whether they are single or multiply banded. Bennetts and Hoskins (1979) suggested that frontal rainbands form when latent-heat release renders the atmosphere symmetrically unstable to saturated ascent (so-called conditional symmetric instability, CSI), apparently supported by some observational studies (e.g., Seltzer et al. 1985; Wolfsberg et al. 1986). Schultz and Schumacher (1999) reviewed the literature and found that even if a region of the atmosphere meets the criteria for CSI, banded precipitation may or may not occur, results since confirmed by other studies (e.g., Schultz and Knox 2007; Schumacher et al. 2010).

The purpose of this paper is to examine the formation and evolution of banded precipitation within idealized baroclinic waves to better understand their sensitivity to diabatic processes, specifically surface friction, the release of latent heat, and surface sensible- and latent-heat fluxes. To address these issues, a three-dimensional primitive-equation model is used to simulate precipitation formation, movement, and dissipation of precipitation bands sometimes differ dramatically between high-resolution numerical-weather-prediction model output and observed radar data (e.g., Novak and Colle 2012). Thus, any improvement in understanding the factors that affect their formation and evolution may be helpful in improving weather forecasts. Despite the research described above, what has not been examined to date in the context of three-dimensional idealized baroclinic waves is the sensitivity of the bands to diabatic processes such as surface friction, latent-heat release, and surface sensible- and latent-heat fluxes.

The formation, movement, and dissipation of precipitation bands at 1-km grid spacing, expressed as precipitation rate in mm h$^{-1}$ during the passage of fronts over the British Isles. (a) a single band at 1600 UTC 22 Nov 2012; (b) multiple bands at 0330 UTC 19 Nov 2012. In (a), the leading edge of the wide band over England and Wales corresponds to the analyzed cold front on the Deutscher Wetterdienst chart (http://www.wetter3.de/Archiv/archiv_dwd.html). In (b), the rearmost band, over western Ireland, corresponds to the analyzed cold front on the Deutscher Wetterdienst chart.
bands within an idealized baroclinic wave. Our motivation is to demonstrate how precipitation banding evolves within the cyclone, free from the complications of real-data cases, such as orography, land–sea differences, and case-to-case variability. A control simulation over the ocean without latent-heat release and surface fluxes is compared to other simulations in which the roughness length of the lower boundary is varied, and the release of latent heat, and the surface heat and moisture fluxes are included. Thus, this paper demonstrates how these diabatic processes affect the structure and evolution of precipitation banding in extratropical cyclones.

The remainder of this article is organized as follows. Section 2 details the mesoscale model and initial conditions used to simulate idealized baroclinic waves and explains the modeling strategy, specifically the sensitivity experiments to be performed. The results of the simulations are presented in the next four sections. Section 3 documents the formation of precipitation bands in the control simulation. Section 4 explores the effect that varying surface friction has on the bands. Section 5 examines how the release of latent heat affects the formation of multiple bands. Section 6 examines the effect of varying the magnitude of surface sensible- and latent-heat fluxes on the formation of multiple bands. Section 7 compares our simulations to previously published observations of precipitation bands to understand the generality of our results. Section 8 concludes this article.

2. Model setup

Moist idealized baroclinic-wave simulations were performed with version 3.4.1 of the Advanced Research Weather Research and Forecasting model (ARW–WRF, Skamarock et al. 2008, hereafter just WRF). The model was initialized by the WRF baroclinic-wave test case, which consists of a zonal jet of roughly 20 K (1000 km) at the surface, spanning 8000 km from north to south on an $f$-plane (where $f = 10^{-4}$ s$^{-1}$ is the Coriolis parameter) in thermal wind balance. The jet is obtained by inverting a baroclinically unstable potential-vorticity (PV) distribution in the $y$–$z$ plane, as in Rotunno et al. (1994). The domain is periodic from west to east and simulations were performed with a wavelength of 4000 km, which is the wavelength of the most unstable normal mode of the WRF initial jet (Plougonven and Snyder 2007, whose initial condition this study uses). The focus in this paper is on the organization of bands spaced around 100 km apart (e.g., Fig. 1b), so a 20-km grid spacing was chosen with 80 vertical levels from the surface up to 16 km. The simulations were run for 204 h with a time step of 120 s.

To investigate how bands may form over a zonally homogeneous ocean surface, a control simulation, OCEANFRIC, is chosen with surface friction and microphysics, but the latent heat of condensation, $L$, is set to zero (Table 1). In subsequent simulations, the surface friction is varied and latent-heat release, then surface heat and moisture fluxes, are incrementally switched on (Table 2). In this way, the
Figure 3: Evolution of OCEANFRIC, between 90 h and 144 h, showing, at given times into simulation: (a,c,e,g) Surface precipitation rate (colored, mm h\(^{-1}\)), \(\theta\) (grey contours every 3 K), and wind vectors (m s\(^{-1}\)). Domain re-centered around precipitation distribution, as in all horizontal plots in this paper. (b,d,f,h) Cross section at location indicated by straight line in lefthand panel, showing relative humidity (colored, %), \(w\) (labeled black contours at \(\pm 0.01, \pm 0.02, \pm 0.03, \pm 0.04, \pm 0.05, \pm 0.075,\) and \(\pm 0.1\) m s\(^{-1}\), negative contours dashed), \(\theta\) (grey contours every 3 K), and the dry-PV 2-PVU contour (bold). Wind vectors not drawn in panel g for clarity of precipitation features. Important \(w\) maxima are numbered in cross sections and numbers correspond between panels and to subsequent cross sections (e.g., \(w1\) is the same \(w\) maximum throughout the paper). Note, the low-level PV in panels d,f,h (and subsequent cross sections throughout this paper) is not diabatic, but due to the strong numerical diffusivity in grid boxes in frontal zones above a no-slip boundary (Cooper et al. 1992).

<table>
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<th>Table 1: Parameterizations in OCEANFRIC</th>
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<td>Microphysics</td>
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<td>Surface layer</td>
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<td>Boundary layer</td>
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Additional simulations employed the same parameterizations, but with varying roughness length, \(L \neq 0\) in the microphysics, and surface fluxes included in the surface-layer parameterization, as detailed in Table 2.

The magnitude of surface friction controlled the pre-occlusion stage of the cyclone’s development and formation of its fronts. To showcase this sensitivity to friction, section 3 presents the early evolution of precipitation bands in OCEANFRIC, and section 4 compares NOFRIC, OCEANFRIC, and LANDFRIC, which differ only in roughness length. In the simulations where friction was included, a constant value of roughness length was prescribed across the whole domain, equal to 0.2 mm and 200 mm for ocean (OCEANFRIC) and land (LANDFRIC), respectively, following Hines and Mechoso (1993).

The effects of setting \(L \neq 0\) and including surface fluxes were most evident in the mature stage of the cyclone. The LATENT and FLUX simulations (in which latent heat and surface fluxes are included incrementally from OCEANFRIC) are presented in sections 5 and 6, respectively. In FLUX, the first simulation in which surface heat and moisture fluxes were active, a sea-surface-temperature (SST) distribution was prescribed equal to the initial temperature of the lowest model level, following Adamson et al. (2006). Another simulation, FLUX–10, is presented in section 6, which is equivalent to FLUX, but with a SST distribution with 10 K subtracted from each grid point.

These simulations are not fully representative of extratropical cyclones over ocean and land. In particular, there is no land-surface or radiation parameterization, hence no diurnal effects, which, particularly over land, may be expected to influence cyclones in the real atmosphere. However, the purpose of this paper is not to perform full-physics baroclinic-wave simulations, but rather to isolate the effects on precipitation bands of surface friction (NOFRIC, OCEANFRIC versus LATENT), and surface fluxes (LATENT, FLUX, and FLUX–10).

### 3. Formation of precipitation bands in control simulation

We now look at the pre-occlusion stage of baroclinic-wave development, and the formation of vertical-velocity maxima and precipitation bands. Section 3a gives an overview of OCEANFRIC (the control simulation, which has a level of friction appropriate to the ocean) and the evolution of its bands. Section 3b looks more closely at the mechanisms by which the vertical-velocity maxima and precipitation bands initially form.

<table>
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<th>Table 2: A summary of all simulations</th>
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<td>(z_0) (mm)</td>
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<tr>
<td>OCEANFRIC</td>
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<td>NOFRIC</td>
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<td>FLUX</td>
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<td>FLUX–10</td>
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Gives roughness length \((z_0)\), whether latent heating ("LH") was left active (Y) or turned off (N), the SST distribution (for simulations with this entry blank, surface fluxes were switched off; \(T_0\) is the initial temperature of the lowest model level) and in which section(s) results are presented. In all simulations in which latent heat was switched on, the Kain–Fritsch cumulus scheme (Kain 2004) was also included, which the model requires to be switched off when latent heating is excluded from the microphysics.
is reduced to 2 K, and no relative humidity or PV drawn, instead showing negative sector. Cross sections are righthand panels in Fig. 3, but only extending to 500 hPa, hydrostatic pressure level, respectively. Middle and righthand panels show cross sections across cold and warm fronts, respectively, at locations indicated in lefthand panels by black and red straight lines, respectively, meeting in the warm sector. Cross sections are as righthand panels in Fig. 3, but only extending to 500 hPa, \( \theta \) contours are red and interval is reduced to 2 K, and no relative humidity or PV drawn, instead showing negative \( \partial \theta / \partial z \) (colored, K km\(^{-1}\)), indicating absolute instability.

### a. Overview of control simulation

The low center deepens primarily between about 90 h and 144 h (Fig. 2a: crosses on OCEANFRIC curve), which is also the period during which the surface-\( \theta \) gradient intensifies (Fig. 2b: crosses on OCEANFRIC curve). Fig. 3 shows the evolution of the system during this period.

Initially, a single band of synoptic-scale precipitation lies over the developing surface warm front (Fig. 3a), corresponding to a single wide maximum of vertical velocity throughout the troposphere (Fig. 3b, \( w1 \)). Then, a narrow maximum of vertical velocity, \( w2 \), begins to separate from \( w1 \) (Fig. 3d), originating at the triple point of the cyclone (Fig. 3c). The wide precipitation band subsequently develops equatorward and, although not producing a rainband, \( w2 \) also develops equatorward at the leading edge of the cold advection (Fig. 3e,f). By 132 h, a wide vertical-velocity maximum, \( w3 \), has formed above and behind the surface cold front, but is in the dry air above the cloud head (Fig. 3f) and so is not precipitating (Fig. 3e).

After 144 h, the surface cold front has become well defined at the equatorward end of the wide precipitation band and there is a poorly defined narrow rainband along this part of the cold front (Fig. 3g), through which the cross section passes. The narrow band is where \( w2 \) has become of greater magnitude, more vertically oriented, and more clearly separated from \( w1 \) (Fig. 3h). Although \( w2 \) is well resolved in this simulation, 20-km grid spacing is insufficient to resolve the associated rainband, as will be illustrated in section 4b. This time is the closest that a cold-frontal rainband comes to forming in this simulation.

The above vertical-velocity maxima are similar to those in previous idealized modeling studies. The wide warm-frontal vertical-velocity maximum, \( w1 \), resembles the warm-sector-wide vertical-velocity band\(^1\) simulated by Knight and Hobbs (1988) and Bénard et al. (1992a). The narrow cold-frontal maximum, \( w2 \), resembles the narrow surface-cold-frontal bands of vertical velocity simulated by Knight and Hobbs (1988), Bénard et al. (1992a), and Zhang and Cho (1995). The wide cold-frontal vertical-velocity maximum, \( w3 \) resembles the wide cold-frontal band of vertical velocity simulated by Bénard et al. (1992a).

### b. Initial band formation

We now look at details of OCEANFRIC during this formative period, specifically how the model generates the different vertical-velocity maxima. Closer inspection of the developing fronts, before any of the additional vertical-velocity maxima and precipitation bands in Fig. 3 have formed, shows a weak maximum of surface-to-850-hPa mean lapse rate (hereafter, just “lapse rate”) lying above the developing cold front and warm sector (Fig. 4a). A cross section across the developing cold front demonstrates that the shallow postfrontal boundary layer is absolutely unstable (Fig. 4b). However, at this time, there is descent where there is cold advection (Fig. 4b) and ascent where there is warm advection (Fig. 4c, the \( w1 \) maximum in Fig. 3), as explained by quasi-geostrophic theory (Holton and Hakim 2013, chapter 6). Therefore, the ascent and instability are isolated from one another and so there is no lift to release the instability.

Six hours later, the lapse-rate maximum lies closer to the wide precipitation band than before (Fig. 4d). Thus, the cold- and warm-frontal cross sections collectively reveal that the unstable postfrontal boundary layer and warm-frontal ascent have come closer together, so that a maximum within the ascent, \( w2 \), begins to form near the triple point (Figs. 4e,f). This moment marks the formation of the \( w2 \) maximum shown in Fig. 3, which eventually forms the poorly defined narrow rainband.

Another 12 h later, the lapse-rate maximum has narrowed and intensified as the warm sector becomes narrower (Fig. 4g). The postfrontal boundary layer is well defined and much more unstable than before, so that \( w2 \) has also intensified (Fig. 4h). Furthermore, the wide cold-frontal \( w3 \) maximum, shown in Fig. 3, has formed in conjunction with a mid-level trough (not shown), so that there is large-scale ascent up the cold front, where there was previously large-scale descent (compare Figs. 4e and h).

Thus, the model produces \( w1 \) from synoptic-scale warm advection, then \( w2 \) at the surface cold front as a narrow instability maximum forms near the triple point, within the synoptic-scale ascent, and finally \( w3 \) as a mid-level trough forms above and behind the surface cold front. However, in
OCEANFRIC, these vertical-velocity maxima do not produce well defined multiple precipitation bands. Sections 4, 5, and 6 show that variations in surface friction, and switching on latent-heat release and surface fluxes can result in multiple precipitation bands from these and other vertical-velocity maxima.

4. Effects of surface friction on precipitation bands

NOFRIC and LANDFRIC are now compared to OCEANFRIC, during the period documented in section 3. Surface friction affects the large-scale structure of the baroclinic wave in these simulations, and so Section 4a first documents the differences in the cyclone structure between the simulations, which is required to explain the effects of friction on the precipitation bands, presented in section 4b.

a. Sensitivity of cyclone structure to surface friction

In simulations with greater surface friction, the cyclone deepens more slowly and achieves less maximum depth: 883, 948, and 961 hPa in NOFRIC, OCEANFRIC, and LANDFRIC, respectively (Fig. 2a). The maximum surface- \( \theta \) gradient is also less for simulations with more friction (Fig. 2b), particularly the warm front, which in OCEANFRIC reaches only 19% of the magnitude that it does in NOFRIC, whereas the cold front in OCEANFRIC reaches 38% of its magnitude in NOFRIC (as calculated in Fig. 5). The difference in frontal contrast between simulations of different surface friction is because wind speed in the warm sector reduces as roughness length increases, which further reduces frontogenesis along the warm front through the feedback between synoptic-scale forcing and vertical motion (e.g., Hines and Mechoso 1993; Rotunno et al. 1998). Frontogenesis along the cold front, on the other hand, occurs away from the low center, where Ekman pumping is weaker (Rotunno et al. 1998). Hence, postfrontal wind speed is less affected by surface friction. Therefore, in OCEANFRIC (where surface friction is included), the surface cold front attains a magnitude more than double that of the surface warm front (Fig. 5a), which is not the case in NOFRIC (Fig. 5b).

Figs. 6a,c,e show, as in the simulations of Hines and Mechoso (1993), how the cyclone structure changes from a Shapiro–Keyser-type cyclone (Shapiro and Keyser 1990), characterized by a bent-back warm front and warm seclusion, in NOFRIC to a Norwegian-type cyclone (Bjerknes 1919; Bjerknes and Solberg 1922), characterized by a thermal ridge and dominant cold front, in LANDFRIC. This contrast in frontal structure between simulations reflects how warm advection dominates without surface friction, so that the warm air moves a long way poleward and wraps around the low center. By contrast, when friction is included, the warm front is broader and weaker, so that the cold front becomes strong, relative to the warm front, and lies behind a very narrow warm sector.
Figure 6: Comparison between NOFRIC (at 114 h), OCEANFRIC (at 144 h), and LANDFRIC (at 162 h), when narrow rainband is at its best defined (if it forms at all). Lefthand panels are as in Fig. 3, except without wind vectors, for clarity of precipitation features. Righthand panels are as in Fig. 3, but with additional $w$ contours of 0.15, 0.2, 0.25, and 0.3 m s$^{-1}$ to emphasize enhancement of ascent at the surface with increasing friction. Boxes drawn in lefthand panels show the locations of the plots at 4-km grid spacing in Fig. 7. Simulations are presented at different times to one another to allow for slower evolution of bands with increasing surface friction.

Figure 7: Comparison of nested domains in NORFRIC, OCEANFRIC, and LANDFRIC at 4-km grid spacing, at times and locations indicated by boxes in Fig. 6. Panels show colored surface precipitation rate (mm h$^{-1}$, note different scale to previous figures) and black contours of surface $\theta$, with contour interval increased from previous figures to 5 K, so not to mask the narrow band. Nested domains extend well beyond area shown, have exactly the same physics as parent domains, and are all initialized 6 h prior to time shown.
b. Sensitivity of precipitation bands to surface friction

The consequences of the above differences in cyclone structure on the precipitation bands are as follows. In all simulations, the wide warm-frontal band forms with the synoptic-scale warm advection and extends equatorward as the cyclone begins to occlude (Figs. 6a,c,e). Because, in NOFRIC, the narrow vertical-velocity maximum, \( w_2 \) (as shown in OCEANFRIC in Figs. 3 and 4), is along the bent-back warm front (Figs. 6a,b) and therefore close to the wide warm-frontal precipitation band, it does not form a separate precipitation band (Fig. 6a). By contrast, in simulations with friction, \( w_2 \) develops equatorward, along the surface cold front and forms the narrow precipitation band (Figs. 6c,e), although, at 20-km grid spacing, the narrow band is poorly resolved.

Simulations at 20-km grid spacing are shown here because this paper is concerned with bands spaced on the order of 100 km apart. Therefore, additional features that higher-resolution simulations may resolve are generally not of interest. However, in this instance, 20-km grid spacing is limited in producing a band of interest.\(^2\) Therefore, to resolve the narrow band, nests of 4-km grid spacing were inserted in each simulation at the locations indicated by boxes in Figs. 6a,c,e. In NOFRIC, the precipitation maximum at the triple point remains within the wide warm-frontal band (Fig. 7a). In OCEANFRIC, the narrow band does separate and achieves a much greater precipitation rate than the wide band (Fig. 7b). In LANDFRIC, the two bands are farther apart and the narrow band is more intense (Fig. 7c) than in OCEANFRIC.

Although this greater horizontal distance between the two bands as friction increases between simulations is poorly demonstrated in the 20-km simulations (Figs. 6a,c,e), the 20-km simulations demonstrate that \( w_1 \) and \( w_2 \) are increasingly separated between simulations (Figs. 6b,d,f). In LANDFRIC, there is a large distance between \( w_1 \) and \( w_2 \) and there is a larger separation between their respective humidity maxima than in the other simulations.

The 20-km simulations are also sufficient to demonstrate that the intensification of the narrow band with increasing surface friction between simulations (Fig. 7) is due to the corresponding intensification of \( w_2 \) between simulations (Figs. 6b,d,f). In LANDFRIC, \( w_2 \) has intensified to the extent that a prominent gravity wave has been generated above and behind the surface cold front. However, the gravity wave’s vertical-velocity anomalies are in the dry air above the cloud head and so do not precipitate (Plougonven and Snyder 2007 discuss the sensitivity of inertia–gravity waves in WRF idealized baroclinic-wave simulations).

Thus, although \( w_2 \) initially forms in the same manner in all simulations to that shown for OCEANFRIC in Fig. 4 (not shown for NOFRIC or LANDFRIC), the subsequent intensity and location of its associated precipitation maximum, relative to the wide warm-frontal band, depends on the level of surface friction. In these simulations, surface friction enhances multiple banding and, although not quite resolving the narrow band, the 20-km simulations capture the mechanisms by which this multiple banding is enhanced.

These effects of surface friction are consistent with previous literature. First, the narrow band is of greater magnitude with increasing friction between simulations, despite a weaker surface cold front, because the introduction of surface friction and a boundary layer induces frictional convergence at the surface cold front (e.g., Bond and Fleagle 1985; Knight and Hobbs 1988; Bénard et al. 1992a), which is enhanced when the lower boundary is rougher. Second, surface friction retards the eastward motion of the surface cold front (e.g., Braun et al. 1997, 1999; Doyle 1997) and, thus, the narrow band, while bands driven by winds farther aloft are less affected. There is thus a greater horizontal distance between bands forced at the surface and those forced at mid-levels.

5. Enhancement of warm-sector bands by latent heating

We now focus on the period of OCEANFRIC beyond that shown in Fig. 3, after the low has fully deepened (the time subsequent to that marked by crosses in Fig. 2), and compare it to LATENT (in which latent heating and cooling, hereafter, just “latent heat” is switched on). Section 5a documents the formation of multiple bands along the warm conveyor belt in both simulations, and section 5b documents the dissipation of those bands.

a. Formation of multiple bands

After 162 h in OCEANFRIC, \( w_2 \) is no longer saturated (Fig. 8b, compare to the earlier time in Fig. 6d, where \( w_2 \) is within the moist region). The precipitation is falling ahead of the surface cold front, as in the idealized baroclinic wave of Boutle et al. (2010), and is broadly speaking in the form of a single wide band along the warm conveyor belt (Fig. 8a). This band has evolved from the wide warm-frontal band exhibited in Fig. 3, but is hereafter termed a warm-sector band because there is no distinctive warm front at this stage of the simulation. This warm-sector band has developed equatorward, as shown in Fig. 3, and, as in the simulations of Bénard et al. (1992a), persisted beyond the dissipation of the narrow band (in what they term the “no upright convection stage”). Thus, the cold front has become of the katafront variety (e.g., Bergeron

\(^2\)All other bands documented in this study are qualitatively similar in nests of 4-km grid spacing, inserted at the relevant times and in the relevant locations (not shown).
Figure 8: Comparison between OCEANFRIC and LATENT at mature stage (162 h into simulations). Lefthand panels are as in Fig. 3. Righthand panels are as in Fig. 3 but no PV is drawn and cross section only goes up to 400 hPa in order to emphasize mid-tropospheric ascent. Also plotted in righthand panels are negative moist PV ($\frac{1}{\rho} \zeta \cdot \nabla \theta_e$, light grey shading, where $\rho$ is density, $\zeta$ is the 3-D absolute-vorticity vector, and $\theta_e$ is equivalent saturated potential temperature) and negative $\frac{\partial \theta_e}{\partial z}$ (indicating conditional instability, dark grey shading).
Figure 9: As Fig. 8 but after 192 h (30 h later).
A precipitation maximum is on the east side of the wide warm-sector band (Fig. 8a). A cross section across the warm conveyor belt (Fig. 8b) reveals that $w_1$ (the original wide warm-frontal vertical-velocity maximum) has evolved into a double maximum aloft, $w_1$ and $w_4$. A vertical-velocity and humidity minimum lies between them, and another maximum is beginning to form behind $w_4$.

These two vertical-velocity maxima, $w_1$ and $w_4$, correspond to the poorly defined double precipitation band in Fig. 8a. By contrast, in LATENT, there is a clear separation between two warm-sector precipitation bands (Fig. 8c). The eastern band is where the release of latent heat has made $w_1$ narrower, more upright, and of greater magnitude (compare Figs. 8b and d). The western band corresponds to the vertical-velocity and humidity maximum, $w_4$. In contrast to OCEANFRIC, the vertical-velocity and humidity minimum between $w_1$ and $w_4$ is a maximum of descent, rather than a minimum of ascent (compare Figs. 8b and d).

The enhancement of these positive and negative vertical-velocity anomalies in LATENT is illustrated by a strip of conditional instability in OCEANFRIC at about 800 hPa (Fig. 8b) that does not appear in LATENT (Fig. 8d) because the instability has been released. All negative moist PV within this mid-level ascent in OCEANFRIC is associated with conditional instability. Therefore, the release of CSI does not appear to have been relevant in forming these multiple bands, nor is there such evidence for any other bands simulated in this study being associated with the release of CSI.

The enhancement of vertical-velocity maxima and reduction of their horizontal scale by latent-heat release was described by Thorpe and Emanuel (1985). In LATENT, these effects have enhanced a wave-like perturbation in vertical velocity across the warm conveyor belt, producing multiple precipitation bands. This perturbation is consistent with the warm-sector band of vertical velocity of Lafore et al. (1994), and that of Zhang and Cho (1995), both of which also exhibited a double maximum at a similar simulation time.

Xu (1992) also found that latent-heat release can enhance multiple bands when large-scale moist ascent evolves into substructures. Although these substructures form in OCEANFRIC (i.e., in the absence of latent heat), the substructures are magnified and separated horizontally when latent-heat release is included. Therefore, these simulations show that there is a tendency for the warm conveyor belt to develop multiple precipitation bands, which is enhanced by latent-heat release, and hence the release of conditional instability, in the mature stage of the cyclone.

In NOFRIC, by this stage of the simulation, the bent-back warm front has intensified and reduced to a scale that does not permit multiple banding on the scale of hundreds of kilometers (not shown). Therefore, friction is required, as in OCEANFRIC and LATENT (and LANDFRIC, not shown), for the cyclone to develop sufficiently slowly and for the warm front to remain at sufficiently large a horizontal scale for these multiple bands to evolve.

b. Dissipation of multiple bands

In OCEANFRIC, 30 h later at 192 h, the cyclone (Fig. 2a) and fronts (Fig. 2b) have weakened. The multiple warm-sector bands have evolved into a single band (Fig. 9a). The cross section reveals that the previously discussed wave-like perturbation has decayed so that the vertical velocity within the saturated region has returned to a single maximum, the original $w_1$ (Fig. 9b; the remnant of $w_4$ is marked to illustrate its dissipation).

Figure 10: Time series of domain-maximum surface sensible heat flux (HFX, W m$^{-2}$) and central surface pressure (p, hPa) in FLUX and FLUX –10. The given heat-flux maximum at all times is in the postfrontal air mass, qualitatively similar to Fig. 1 of Sinclair et al. (2010).

The addition of latent heat delays the return to a single band, although there is little difference in the magnitude of the low center or fronts at this time (not shown). The double vertical-velocity and humidity maximum are still well defined in LATENT (Fig. 9d) and, thus, there is still a remnant of the double precipitation band (Fig. 9c). As in Fig. 8, the presence of conditional instability co-located with the mid-level ascent in OCEANFRIC (Fig. 9b), but not in LATENT (Fig. 9d), indicates that conditional instability is still being released in LATENT, so that latent-heat release continues to enhance the vertical-velocity maxima. Therefore, in these simulations, latent-heat release enhances multiple bands in the warm sector and increases their longevity.
Figure 11: Comparison at mature stage (162 h into simulations, same time as Fig. 8) between LATENT, FLUX–10, and FLUX. Left-hand panels are as in Fig. 8. Right-hand panels are as in Fig. 8, but without shaded instabilities and with the dry-PV 2-PVU contour in bold to emphasize increasing distance between tropopause fold and w2 with increasing fluxes.
6. Enhancement of anafront by surface fluxes

This section looks at the effects of adding surface heat and moisture fluxes to LATENT. We focus on the same period of the simulations as that in section 5. Section 6a looks at FLUX, in which a SST distribution is prescribed equal to the initial temperature of the lowest model level, and compares it to LATENT. Section 6b looks at the effect of varying the SST.

a. Addition of surface fluxes

In FLUX, because a meridionally varying SST distribution is prescribed, increasing baroclinicity leads to increasing heat flux into the ocean in the warm-sector (as in, e.g., Sinclair et al. 2010). As long as the low remains deep, this postfrontal warm flux remains large (Fig. 10: FLUX curves). After 162 h, \(w^2\) is still intense (Fig. 11f) at a time at which it has considerably weakened in LATENT (Fig. 11b). Thus, unlike LATENT, the warm-sector precipitation is still tied to the surface cold front and intense there, more closely resembling the classic ana-cold-front model (compare Figs. 11a and e).

As further illustration of the ana-kata distinction, in LATENT, the dry intrusion is readily evident in the precipitation field (Fig. 11a), resembling the post-occlusion conceptual model of Schultz and Vaughan (2011, their Fig. 14b). The cross section passes through the dry intrusion, showing that the dry stable air associated with the tropopause fold overlies the considerably weakened \(w^2\) (Fig. 11b). By contrast, in FLUX, the dry slot is less distinctive (Fig. 11c) and the postfrontal boundary layer is much deeper (Fig. 11f), due to the postfrontal heat flux from the ocean into the atmosphere. Therefore, the tropopause fold does not protrude as low and is well behind \(w^2\).

The effect of the ana–kata distinction on the precipitation bands is that \(w^1\) and \(w^2\) have come much closer together with surface fluxes (compare Figs. 11b and f). Therefore, although the double vertical-velocity maxima aloft, \(w^1\) and \(w^4\), do form, as in LATENT, they are underlain by the intense \(w^2\). Therefore, the precipitation has not divided into the double band that forms in LATENT and there is just a single wide band of intense precipitation along and ahead of the surface cold front (Fig. 11e). Thus, the effect of latent heat to produce two separate warm-sector bands (Fig. 8) is reversed by surface fluxes.

b. Sensitivity to sea-surface temperature

To test the sensitivity of the bands to SST, a further simulation, FLUX–10, was performed, equivalent to FLUX, but with SST everywhere and uniformly 10 K lower. There is very little difference between FLUX and FLUX–10 in the deepening rate of the cyclone, but, in FLUX–10, the maximum surface sensible heat flux is 30% less than in FLUX (Fig. 10). This contrast is a result of the same air encountering a cooler sea surface at a given grid point in FLUX–10 than in FLUX. Therefore, FLUX–10 can be considered an intermediary between LATENT and FLUX.

FLUX–10 is shown at the same time as LATENT and FLUX in Fig. 11c,d. FLUX–10 bears qualities of each of the other two. The \(w^2\) maximum is intense and saturated, as in FLUX, but farther behind the warm-conveyor-belt ascent than in FLUX (compare Figs. 11d and f), allowing more space between \(w^1\) and \(w^4\). Thus, the double precipitation band has formed (Fig. 11c), albeit less markedly than in LATENT (Fig. 11a).

Prescribing higher SSTs does not completely remove multiple banding, however. In FLUX, 42 h later at 204 h, unlike the other simulations, the precipitation behind the surface cold front has become organized into a number of bands (compare Figs. 12a,c,e) of postfrontal boundary-layer convection (e.g., Kuettner 1959; Miura 1986; Atkinson and Zhang 1996), due to enhancement of the postfrontal vertical-velocity and humidity anomalies (compare Figs. 12b,d,f). Therefore, although greater SST inhibits multiple warm-sector bands, it enhances multiple postfrontal bands. This later time of the simulations further illustrates the change from kata-type to ana-type cold front as surface fluxes increase between simulations, with the warm-sector precipitation a long way ahead of the surface cold front in LATENT (Fig. 12a), but still tied to the surface cold front in FLUX (Fig. 12e).

For simplicity, we have used varying SST in this work as a proxy for varying surface fluxes. Nonlinearities between the surface layer and lower troposphere mean that real surface fluxes are a complex function of many variables, so our FLUX and FLUX–10 simulations should be taken as a sensitivity test of the model rather than a representation of reality. We include FLUX–10 to illustrate that multiple banding is not precluded by the inclusion of surface fluxes, and that postfrontal bands may not always be present in such simulations.

7. Comparison of simulated bands to observations

The similarity of the bands simulated in this study to other idealized-modeling studies has been noted throughout the text. We now discuss how closely these bands resemble those documented in observational studies.

The wide warm-frontal band, the first band to form in all simulations, is a well-documented aspect of extratropical cyclones. As noted by Houze and Hobbs (1982, p. 234), this band forms from “stratiform precipitation produced by widespread lifting associated with warm advection in the leading portion of the cyclone system”, which is exactly how it formed in these simulations.
Figure 12: As Fig. 11 but after 204 h (42 h later) and 2-PVU contour is omitted for clarity in cross sections.
The narrow band, documented in OCEANFRIC and LANDFRIC, results from boundary-layer convergence and low static stability above the surface cold front, when the cold front is of an ana-type (e.g., Browning and Pardoe 1973; Houze et al. 1976; Hobbs and Biswas 1979; James and Browning 1979; Browning 1986), which is the manner in which the band formed in the simulations with surface friction. Accordingly, the band dissipated in the mature stage of these simulations when the tropopause fold came around the cyclone and over-ran the band (i.e., when the front changed to kata-type).

In the mature stage of the cyclone’s evolution, the wide warm-frontal band evolved into multiple warm-sector bands (when latent heat was included). They are termed as such in this paper because there was no distinctive warm front at this stage of the simulations. Such bands have been reported in previous studies. For example, Browning and Harrold (1969, p. 298) observed, from rain gauges, “bands consisting of small rain areas aligned parallel to the warm sector winds” (illustrated in their Fig. 5d). Browning (1986, p. 31) then observed on radar, “organized bands of moderate to heavy rain several tens of kilometers wide... associated with mesoscale circulations within the warm conveyor belt about an axis parallel to the relative mean flow” (illustrated in his Fig. 7a). The warm-sector bands in these studies were associated with mature cyclones, as in this paper. The two parallel bands in Fig. 1b are also qualitatively similar to our warm sector bands, lying between the analyzed cold and warm fronts.

The postfrontal bands simulated when surface fluxes were included are also easy to relate to observations. Over open ocean, cold air flowing over warm water generates shallow convection, which is organized into bands, parallel to the vertical shear of the horizontal wind, when that shear is sufficient (e.g., Miura 1986). These postfrontal bands are also exhibited in Fig. 1a, in which the spacing and orientation of the postfrontal bands, relative to the principal cold-frontal band, closely resemble those in Fig. 12c.

Although all the bands simulated in these experiments are related to similar ones described in the literature, wide cold-frontal bands and wavelike bands (Houze et al. 1976) and prefrontal cold-surge bands (Hobbs 1978) were not simulated, neither were the northwest bands in Novak et al. (2004). We suggest that the explanation may be a result of excluding large-scale deformation in the background flow (e.g., Schultz et al. 1998; Schultz and Zhang 2007) and nonuniform lower-boundary conditions (e.g., Physick 1988; Doyle 1997; Muir and Reeder 2010). Addressing these hypotheses would require further experiments, which are outside the scope of this study.

8. Summary

Idealized baroclinic waves were simulated at 20-km grid spacing and the evolution of the associated precipitation bands documented. This study follows up studies of multiple vertical-velocity banding in two-dimensional (e.g., Knight and Hobbs 1988; Bénard et al. 1992a,b; Xu 1992; Lafore et al. 1994; Pizzamei et al. 2005) and three-dimensional (e.g., Zhang and Cho 1995; Gray and Daacre 2008) idealized models. The new elements are the focus on the formation and structure of precipitation bands in a three-dimensional model, and the investigation of sensitivity to a greater variety of physical processes (surface friction, latent-heat release, and surface sensible- and latent-heat fluxes).

In all simulations, initially, the vertical-velocity and hence precipitation distribution was driven by synoptic-scale warm advection, leading to a single wide warm-frontal precipitation band. As the surface fronts began to develop, a lower-tropospheric lapse-rate maximum developed near the triple point, producing a narrow maximum of vertical velocity within the synoptic-scale ascent.

When the lower boundary was frictionless, the low-pressure center deepened and warm advection intensified more than for a rough lower boundary. A Shapiro–Keyser-type cyclone was thus produced with an intense bent-back warm front. The precipitation thus became concentrated along the bent-back front and, although a narrow precipitation maximum formed along the bent-back front, due to the narrow vertical-velocity maximum, the precipitation maximum did not become a separate band. Therefore, multiple bands did not form in the frictionless case.

As friction was increased to that of an ocean surface in a different simulation, warm advection was reduced and the cold front became stronger, relative to the warm front. The narrow vertical-velocity maximum, which in the frictionless case formed along the bent-back front, in this simulation formed along the surface cold front and developed equatorward along the surface cold front. Within a nested domain of 4-km grid spacing, this vertical-velocity maximum formed a narrow rainband, separate from the wide warm-frontal band, in contrast to the frictionless case, where the narrow precipitation maximum remained within the warm-frontal band.

As friction was increased to that of a land surface in a further simulation, a Norwegian-type cyclone evolved, with the cold front even stronger, relative to the warm front. The narrow rainband in the nested domain was more intense, due to the greater surface friction, and a greater horizontal distance formed between the narrow band and the wide warm-frontal band. Thus, the effect of friction in these simulations was to separate horizontally the two precipitation bands and enhance the narrow band.

In the mature stage of the ocean-friction case, in the absence of surface fluxes, the surface cold front became
over-run by the tropopause fold. The narrow rainband, whose associated ascent was at the surface cold front, thus dissipated and, subsequently, the precipitation distribution became centered on the warm conveyor belt. In the absence of latent-heat release, a wave-like perturbation in vertical velocity and humidity formed across the warm conveyor belt at this stage, but did not produce multiple precipitation bands. With the inclusion of latent-heat release, this perturbation became much more pronounced, so that the vertical-velocity maxima were narrower, more upright, and of greater magnitude, and well defined downdrafts formed between updrafts. Multiple precipitation bands thus developed in the warm-sector. These bands persisted until the cyclone’s low center began to decay, at which point the vertical-velocity and humidity perturbations also decayed and the multiple bands dissipated.

When surface fluxes were included, with the SST equal to the initial temperature of the lowest model level, this warm-sector precipitation remained tied to an ana-type cold front, the cold front moving at a greater rate eastward than without surface fluxes. The postfrontal boundary layer was much deeper, and the tropopause fold did not protrude so low and stayed well behind the surface cold front. Thus, the ascent associated with the narrow rainband remained intense throughout this mature stage of the cyclone. Although multiple vertical-velocity maxima were generated aloft, as in the absence of surface fluxes, they were underlain by the still-intense ascent at the surface cold front. Therefore, a single intense precipitation band lay along and ahead of the surface cold front, to which both the surface-based and mid-tropospheric ascent contributed. However, finer-scale postfrontal bands were generated during this stage of the cyclone’s lifecycle that did not form in the absence of surface fluxes. When the SST was decreased by 10 K, the simulation more closely resembled the no-surface-fluxes case: multiple warm-sector bands did form and postfrontal bands did not form. Therefore, the inclusion of surface fluxes in baroclinic-wave simulations does not preclude the multiple warm-sector bands, nor make the postfrontal bands inevitable.

These simulations show the relative importance of surface friction, latent-heat release, and surface fluxes on the distribution of precipitation within extratropical cyclones. Further idealized-modeling studies with different initial and lower-boundary conditions are needed to show banding tendencies for more specific cyclone types and in specific geographic locations.

Acknowledgements

The precipitation-radar data plotted in Fig. 1 were provided by the Met Office through the British Atmospheric Data Centre (BADC). Thanks to Riwal Plougonven, who provided the code to produce the initial jet, and Tim Slater, who altered the code to produce a jet at 20-km grid spacing. We also thank the two anonymous reviewers whose comments have improved the manuscript. Jesse Norris is a NERC-funded student through the DIAMET (DIAbatic influences on Mesoscale structures in ExTratropical storms) project, NE/I005234/1.
REFERENCES


4 Paper 2: Precipitation cores along a narrow cold-frontal rainband in idealized baroclinic waves.
Precipitation cores along a narrow cold-frontal rainband in idealized baroclinic waves

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ABSTRACT
Precipitation cores (PCs) often form along a narrow cold-frontal rainband, where cores of maximum precipitation rate, rotated clockwise from the orientation of the cold front, are separated by gaps of lighter or no precipitation in between, as observed by precipitation-radar data. Kawashima (2011) showed that variations in the width, length, intensity, and orientation of precipitation structures that form from horizontal shear instability, resembling observed PCs, are due to variations in the cross-frontal horizontal shear, cross-frontal vertical shear, and static stability in the cold-frontal environment. This study shows in a more realistic idealized baroclinic-wave model, first in a control simulation, that these variations in shear and static stability may exist along the length of the cold front, and hence that a wide variety of PCs may exist simultaneously, which is also exhibited by radar data. Second, sensitivity simulations are performed, in which diabatic influences (latent-heat release, surface friction, and surface sensible- and latent-heat fluxes) are varied between simulations. These simulations demonstrate how variations in these diabatic influences lead to the variations in shear and static stability that may affect PCs in the real atmosphere. Latent-heat release and surface friction enhance PCs, whereas larger surface fluxes lead to a more uniform cold-frontal rainband. These simulations may explain some of the variability of PCs that are observed, both along a cold front in a single case and between different cases.

1. Introduction
The distribution of precipitation along a surface cold front is often observed to become concentrated into precipitation cores (PCs) of clockwise orientation relative to the surface cold front, with gaps of lighter or no precipitation between them (Hobbs and Biswas 1979; James and Browning 1979). PCs are most easily observed over land, due to the availability of precipitation-radar data, e.g., over the British Isles (Fig. 1). However, PCs have also been observed over the ocean (e.g., Hobbs and Biswas 1979; Hobbs and Persson 1982; Wakimoto and Bosart 2000; Jorgensen et al. 2003).

A considerable variety of PCs exists from case to case (Fig. 1). PCs may be long and spaced far apart (Fig. 1a) or short and spaced much more closely together (Fig. 1b). PCs may also be more curved in some cases than others (Figs. 1c,d). Furthermore, PCs may vary within a single case, with the length, width, spacing, curvature, and cold-front-relative angle of PCs varying along the cold front (particularly exhibited in the boxes drawn in Figs. 1a,c). PCs may even be embedded within a wide band along the cold front (Fig. 1d). Given this wide range of possible morphologies, any research examining the physics associated with PCs may be helpful in gaining a better understanding of why different PCs vary as observed.

PCs have commonly been proposed to result from horizontal shear instability, for which many studies have gained observational evidence (e.g., Matejka et al. 1980; Carbone 1982; Hobbs and Persson 1982; Browning and Roberts 1996). Also, some studies have performed real-data simulations of PCs, in which the simulated wind field has properties that agree well with horizontal-shear-instability theory (e.g., Brown et al. 1999; Jorgensen et al. 2003; Smart and Browning 2009). Perhaps most compellingly, Kawashima (2011) used an idealized shear-deformation model to simulate precipitation structures along a cold front from horizontal shear instability, resembling observed PCs. However, the same author simulated, in another modeling study, realistic-looking idealised PCs forming from a wind field that did not agree with horizontal-shear-instability theory (Kawashima 2007). Furthermore, Brown et al. (1999), whose real-data simulation of observed PCs exhibited properties agreeing with horizontal-shear-instability theory, also found key differences. As noted by Brown et al., “...shear instability has not been demonstrated to be a dominant mechanism for alongfront variability in realistic cold fronts with moist ascent...”, which, to our knowledge, remains the
This study is less interested in the theory of PC formation and more interested in what differences in the flow and stratification in the vicinity of the cold front lead to such differences in the morphology of PCs from case to case, but also along the cold front within a single case, as exhibited in Fig. 1 and described above. For this reason, we are particularly motivated by Kawashima (2011), whose idealized-modeling study investigated the sensitivity of PCs to:

- the vertical shear of the horizontal wind,
- the cross-frontal shear of the horizontal wind, and
- the prefrontal static stability.

Increasing the cross-frontal vertical shear, relative to the along-frontal vertical shear, between simulations led to greater amplitude and growth rate of the perturbation with which the PCs are associated, hence PCs oriented at a greater angle to the cold front and larger gaps along the cold front between PCs (his Figs. 7 and 9). On the other hand, increasing the cross-frontal shear of the horizontal wind between simulations led to a greater initial wavelength, hence longer PCs and larger gaps (his Fig. 20). Decreasing the static stability between simulations led to more poorly organized PCs, which, when vertical shear was high, implied bow-shaped PCs with great curvature, and, when vertical shear was low, short PCs resembling cellular convection (his Figs. 7 and 9, compare rows of panels).

Thus, in Fig. 1, the PCs at large angles to the cold front may be associated with strong cross-frontal, relative to along-frontal, vertical shear. The long and short PCs in Fig. 1 may be associated with strong and weak wind shifts, respectively, across the cold front. The less distinctive, but intense, PCs may be associated with large near-surface lapse rates. The PCs that Kawashima (2011) simulated formed from horizontal shear instability, so their sensitivity to these factors may or may not be the same as for PCs in the real atmosphere. Either way, Kawashima (2011) provides testable hypotheses of PC sensitivity in a more realistic model.

The approach of this paper is to use a three-dimensional primitive-equation model to simulate PCs along a cold front within an idealized baroclinic wave. As such, we cannot easily adjust the shear and lapse rate between simulations, but we can do so indirectly by varying the model physics, and by changing the thermal and frictional properties of the lower boundary. Thus, we can learn what influences on a precipitating cold front (e.g., latent-heat rate, surface friction, and surface sensible- and latent-heat fluxes) may lead to the given variations in shear and static stability and, thus, how sensitive PCs are to these influences (if indeed the PCs in a primitive-equation model are sensitive to shear and static stability in the same ways as in a simpler shear–deformation model).

As in Norris et al. (2013), hereafter N13, different simulations are compared in which latent-heat release, roughness length, and sea-surface temperature (SST) are varied. A simulation from N13 with all these diabatic factors appropriate to an extratropical cyclone over the open ocean, after 132 hours when the surface cold front has formed, is used as an initial condition for all simulations and the different diabatic factors are varied between simulations, subsequent to this time. Thus, this paper demonstrates how PCs along a mature cold front of maritime origin vary, depending on differences in latent-heat release, surface friction, and surface fluxes that the cold front encounters.

The remainder of this paper is organized as follows. Section 2 describes the model and initial condition used, and distinguishes the control simulation from the various sensitivity simulations performed. Section 3 documents the evolution of PCs along the cold front in the control simulation and shows how the PCs vary along the cold front. With the key characteristics of the control simulation established, the sensitivity simulations are then presented as comparisons to the control simulation. These sensitivity simulations are performed for latent-heat release, surface friction, and surface fluxes separately, which are presented in sections 4, 5, and 6, respectively. Finally, section 7 concludes this paper.

2. Initial condition and methodology

Moist idealized baroclinic-wave simulations were performed with version 3.4.1 of the Advanced Research Weather Research and Forecasting model (ARW–WRF, Skamarock et al. 2008, hereafter just WRF). As in N13, the model was initialized by the WRF baroclinic-wave test case, which consists of a zonal jet of roughly 20 K (1000 km)$^{-1}$ at the surface, spanning 8000 km from north to south on an f-plane ($f = 10^{-4}$ s$^{-1}$) in thermal wind balance. The jet is obtained by inverting a baroclinically unstable potential-vorticity distribution in the $y-z$ plane, as in Rotunno et al. (1994). The domain is periodic in the $x$-direction, with a wavelength of 4000 km, which is the wavelength of the most unstable normal mode of the WRF initial jet (Plougonven
Sfc. precip. (colored), $\theta$ (grey), winds.

Cross-section $w$ (black), $\theta$ (red).
Fig. 2. The initial condition for all simulations in this paper, showing an idealized baroclinic wave at 20-km grid spacing, 132 h after an initial perturbation is made to a uniform midlatitude jet, as described in N13. In the 132 h that the baroclinic wave has evolved to this state, the model has included the parameterizations and surface-layer specification detailed in Table 1. Panels show: (a) Surface precipitation rate (colored, mm h\(^{-1}\)), \(\theta\) (grey contours every 3 K), wind vectors (m s\(^{-1}\)), the location of the nested domain (which is initialized at this time) indicated by the black box, and the location of the coastline (which is created at this time in all simulations other than CNTL and NOLATENT) indicated by the bold red line. (b) Cross-section at location indicated by bold black line in panel a, showing \(\theta\) (red contours every 3 K), and \(w\) (black contours at \(\pm 0.01, \pm 0.02, \pm 0.03, \pm 0.04, \pm 0.05, \pm 0.075\), and \(\pm 0.1\) m s\(^{-1}\), negative contours dashed). The annotated \(w_1\) and \(w_2\) vertical-velocity maxima are for reference to N13.

The FLUX simulation in N13 (whose specifics are given in Table 1) uses this initial condition and simulates a baroclinic wave at 20-km grid spacing, with 80 vertical levels from the surface up to 16 km, with diabatic processes appropriate to an extratropical cyclone above the open ocean. After 132 h, the initial condition for all simulations in this paper, a wide band of precipitation lies along and ahead of a well-defined surface cold front and wind shift (Fig. 2a). This precipitation is due partly to the wide vertical-velocity maximum above the surface warm front, \(w_1\), and partly to the narrow vertical-velocity maximum at the surface cold front, \(w_2\) (Fig. 2b). The evolution of both of these maxima to their current state was described in N13. This paper is concerned with \(w_2\) only and the variation of its associated rainfall along the cold front.

While N13 was concerned with precipitation bands spaced several hundreds of km apart and, thus, used a 20-km grid spacing, this study is concerned with PCs spaced several tens of km apart. Therefore, a nested domain of 4-km grid spacing is inserted at the location indicated by the black box in Fig. 2a and the passage of the cold front across the nest is documented over the 36 h that the cold front takes to cross the nest.

Table 1. Specifications of FLUX simulation in N13 that produces this paper’s initial condition

<table>
<thead>
<tr>
<th>Microphysics</th>
<th>Thompson (Thompson et al. 2008)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Surface layer</td>
<td>MM5 Monin-Obukhov (Monin and Obukhov 1954)</td>
</tr>
<tr>
<td>Boundary layer</td>
<td>Yonsei University (Hong et al. 2006).</td>
</tr>
<tr>
<td>Convection</td>
<td>Kain-Fritsch (Kain 2004)</td>
</tr>
<tr>
<td>(z_0)</td>
<td>0.2 mm</td>
</tr>
<tr>
<td>SST</td>
<td>Initial (T) of lowest model level</td>
</tr>
</tbody>
</table>

Gives all parameterizations used (top four entries) and the roughness length \((z_0)\) and sea-surface-temperature distribution (SST).

In the control simulation (CNTL), the simulation is kept the same as in Table 1. In the other simulations, the lower boundary is converted to half-ocean–half-land, with ocean and land occupying the lefthand and righthand side of the outer domain, respectively (the bold red line in Fig. 2a indicates the location of the coastline). Therefore, the nested domain is almost all land in the sensitivity simulations and the coverage of these simulations in this paper is of the movement of the surface cold front over the land. However, the nested domain also contains 100 grid boxes in the \(x\)-direction over the ocean, west of the coastline, in order that the land–sea contrast is resolved in the nest.

In all simulations, the lefthand ocean side of the domain has the same constant roughness length of 0.2 mm and a SST distribution equal to the initial temperature of the lowest model level (when the baroclinic wave is initialized, as opposed to at the initialization time in this paper), \(T_0\). The variability in the sensitivity simulations is all in the thermal and frictional characteristics of the righthand land side of the domain, as summarized in Table 2, except in NOLATENT, in which latent-heat release is subtracted from the microphysics (which can only be done in the model if it is switched off everywhere, over both domains). The LANDFRIC1, LANDFRIC2, and LANDFRIC3 simulations contain a frictional contrast only between land and Snyder 2007).

Table 2. A summary of all simulations

<table>
<thead>
<tr>
<th>(z_0) (mm)</th>
<th>LH</th>
<th>SST (K)</th>
<th>Sections</th>
</tr>
</thead>
<tbody>
<tr>
<td>CNTL</td>
<td>0.2</td>
<td>Y</td>
<td>(T_0) 3,4,5,6</td>
</tr>
<tr>
<td>NOLATENT</td>
<td>0.2</td>
<td>N</td>
<td>(T_0) 4</td>
</tr>
<tr>
<td>LANDFRIC1</td>
<td>5</td>
<td>Y</td>
<td>(T_0) 5</td>
</tr>
<tr>
<td>LANDFRIC2</td>
<td>250</td>
<td>Y</td>
<td>(T_0) 5</td>
</tr>
<tr>
<td>LANDFRIC3</td>
<td>2000</td>
<td>Y</td>
<td>(T_0) 5</td>
</tr>
<tr>
<td>MINUS5K</td>
<td>0.2</td>
<td>Y</td>
<td>(T_0 - 5) 6</td>
</tr>
<tr>
<td>MINUS2K</td>
<td>0.2</td>
<td>Y</td>
<td>(T_0 - 2) 6</td>
</tr>
<tr>
<td>PLUS2K</td>
<td>0.2</td>
<td>Y</td>
<td>(T_0 + 2) 6</td>
</tr>
<tr>
<td>PLUS5K</td>
<td>0.2</td>
<td>Y</td>
<td>(T_0 + 5) 6</td>
</tr>
</tbody>
</table>

Gives roughness length \((z_0)\), whether latent heating ("LH") was left active (Y) or turned off (N), the SST distribution \((T_0)\) is the initial temperature of the lowest model level when the baroclinic wave is initialized) and in which section(s) results are presented. Bold entries are the given factor is different to CNTL. The given factors are prescribed on the righthand side of the domain only (to the right of the red line in Fig. 2) and, in all simulations, the lefthandside of the domain is as in CNTL, except in NOLATENT because the model can only switch off latent-heat release if over the entire domain.
Fig. 3. Surface precipitation rate (colored, mm h$^{-1}$) and $\theta$ (black contours every 3 K) in part of the nested domain in CNTL after 18 h. “Poleward” and “equatorward” boxes show locations of finer-scale plots presented in subsequent figures as indicated. In all panels in this paper referred to as the poleward and equatorward boxes, the range of $x$ co-ordinates varies from panel to panel, depending on the location of the cold front at the given time of the given simulation, but the range of $y$ co-ordinates is always as indicated in this figure (i.e., the same part of the cold front). Smaller red box shows location of even finer-scale plot in Fig. 5a.
Sfc. precip. (colored), $\theta$ (grey), winds. Sfc. rel. vort. (black), $\theta$ (red).
Fig. 4. Evolution of CNTL in equatorward box, whose location is shown in Fig. 3, at indicated times. Lefthand panels show surface precipitation rate (colored, mm h$^{-1}$), $\theta$ (grey contours every 2 K), and wind vectors (m s$^{-1}$). Righthand panels show the vertical component of relative vorticity at the surface $\left(\frac{\partial v}{\partial x} - \frac{\partial u}{\partial y}\right)$ (black contours at 5, 10, 20, 50, and $100 \times 10^{-5}$s$^{-1}$) and surface $\theta$ (red contours every 2 K).

Sfc. precip. (colored), $\theta$ (red), winds. Cross-section $w$ (black), $\theta$ (red).

Fig. 5. (a) Finer-scale plot of CNTL after 18 h of a single PC at location indicated by red box in Fig. 3, showing surface precipitation rate (colored, mm h$^{-1}$), $\theta$ (red contours every 2 K), and wind vectors (every grid box, m s$^{-1}$). (b,c,d) Cross sections up to 500 hPa at locations indicated by black, red, and blue lines in (a), showing $w$ (black contours at $\pm0.01$, $\pm0.02$, $\pm0.03$, $\pm0.04$, $\pm0.05$, and $\pm0.1$ m s$^{-1}$, negative contours dashed) and $\theta$ (red contours every 2 K). The $w2$ vertical-velocity maximum is annotated for reference to Fig. 2.
and sea, with respective land roughness lengths of 5 mm, 250 mm, and 2000 mm prescribed (appropriate, respectively, for featureless land, high crops, and a city center, World Meteorological Organization 2008). The MINUS5K, MINUS2K, PLUS2K, and PLUS5K simulations contain a thermal contrast only, with respective land skin temperatures of $T_0 - 5 K$, $T_0 - 2 K$, $T_0 + 2 K$, and $T_0 + 5 K$. However, the term “land” is used loosely and in no simulations is a land-surface scheme used. The purpose of this paper is not to simulate a full physics land–sea contrast. Instead, in LANDFRIC1, LANDFRIC2, LANDFRIC3, MINUS5K, MINUS2K, PLUS2K, and PLUS5K, the lower boundary is treated as an ocean surface with a roughness-length or SST discontinuity, in order to isolate the sensitivity of PCs to each of these factors. “Coastline” is used to refer to the discontinuity, whether frictional or thermal.

The variations between simulations detailed in Table 2 are effective only after the simulations are re-initialized at 132 h. The alternative would be to vary the simulations during the full baroclinic-wave lifecycle. However, as shown in N13, varying friction, latent-heat release, and surface fluxes between baroclinic-wave simulations leads to profound differences in the structure and rate of development of the cyclone and fronts. This paper aims to investigate the sensitivity of a given cold front to the various diabatic processes; if these diabatic processes were varied between simulations for the entire baroclinic-wave lifecycle, then the simulations would not be comparing the same cold front. Thus, this paper applies the methodology of N13, but taking a mature cold front, rather than an unperturbed jet, as the initial condition.

3. Precipitation cores in control simulation

This section documents PCs forming along the cold front in CNTL. Our interest is in the evolution of the PCs, their variability along the cold front, and the near-surface shear and static stability associated with them.

Eighteen hours after the nest is initialized (18 h after Fig. 2), about twenty PCs have formed along the surface cold front (Fig. 3). In the “equatorward” box, the PCs are wide, at a large angle to the cold front, and with large gaps of lighter or no precipitation between cores. In the “poleward” box, the PCs are narrower, at a smaller angle to the cold front, and with smaller gaps between cores. This paper predominantly documents PCs in the equatorward box because this is where the PCs are most distinctive and this is where they vary the most between simulations. Section 3a shows the evolution of PCs in the equatorward box, section 3b more closely examines a single PC, and section 3c compares the PCs in the equatorward and poleward boxes.

a. Evolution of precipitation cores

In the equatorward box, the PCs form after about 15 h, at which time they resemble convective cells (Fig. 4a), in conjunction with a noisy wave-like feature in vorticity immediately ahead of the cold front (Fig. 4b). Six hours later, the prefrontal vorticity distribution has evolved into a smoother wave-like feature, with vorticity maxima oriented roughly perpendicular to the cold front (Fig. 4d) and moving poleward along the cold front (not shown). By this time, the clockwise-oriented PCs have become well-defined, evenly spaced, and with distinctive gaps between cores (Fig. 4c). The PCs are subsequently near steady-state for the remainder of the simulation (Figs. 4e,g). However, during this time, the prefrontal vorticity maxima become more aligned along the cold front, possibly in phase with another wave-like disturbance behind the cold front (Figs. 4f,h), with vorticity maxima moving equatorward (not shown), the opposite direction to the prefrontal maxima, and also pointing in the opposite direction to the prefrontal maxima. During this time, the gaps between PCs slightly shrink and the PCs become aligned more closely to the cold front.

Thus, the PCs are associated with counter-propagating wave-like vorticity maxima either side of the surface cold front. Counter-propagating vorticity edge waves, forming on opposite-signed vorticity gradients (such as there are either side of the cold front) can lead to the phase-locking and mutual amplification of the edge waves (e.g., Bretherton 1966). Kawashima (2011) found his idealised PCs to and mutual amplification of the edge waves (e.g., Bretherton 1966). Kawashima (2011) found his idealised PCs to and mutual amplification of the edge waves (e.g., Bretherton 1966). Kawashima (2011) found his idealised PCs to
Sfc. precip. (colored), θ (grey), winds. Low. trop. lapse rate (greys), sfc. θ (red), 950-hPa vert. shear (vectors).

Fig. 6. Comparison of poleward and equatorward boxes (as indicated in Fig. 3) in CNTL after 18 h. Lefthand panels are as in Fig. 4. Righthand panels show mean surface-to-850-hPa lapse rate (grey shading, K km$^{-1}$, calculated by $(T_{\text{surf}} - T_{850}) / (z_{850} - z_{\text{surf}})$, where $T$ and $z$ are temperature and height, and the $\text{surf}$ and $850$ subscripts denote the lowest model level and the 850-hPa hydrostatic-pressure level), surface θ (red contours every 2 K), and vectors of the vertical shear of the horizontal wind at 950 hPa (m s$^{-1}$km$^{-1}$).
Sfc. precip. (colored), $\theta$ (grey), winds. Low. trop. lapse rate (greys), sfc. $\theta$ (red), 950-hPa vert. shear (vectors).

Fig. 7. As Fig. 6 after 30 h (12 h later).
Sfc. precip. (colored), $\theta$ (grey), winds. Low. trop. lapse rate (greys), sfc. $\theta$ (red), 950-hPa vert. shear (vectors).

Fig. 8. As Fig. 6, but comparing equatorward boxes of CNTL and NOLATENT after 18 h.
Sfc. precip. (colored), $\theta$ (grey), winds. Low. trop. lapse rate (greys), sfc. $\theta$ (red), 950-hPa vert. shear (vectors).

Fig. 9. As Fig. 8, but after 30 h (12 h later).
but a very shallow one, extending only to about 950 hPa (Fig. 5d). Thus, there is ascent along the full length of the core and gap, which extends higher and tilts to the right where the wind shift is farthest ahead of the surface cold front.

c. Variation of precipitation cores along cold front

The PCs at the poleward and equatorward ends of the cold front, as indicated by the two boxes in Fig. 3, are now compared. Figs. 6a,c show more clearly the contrast between the two ends of the cold front after 18 h. At the poleward end, the PCs are at less of an angle to the cold front, so that the gaps are smaller and the precipitation distribution more closely resembles a continuous cold-frontal rainband. The wavelength of the PCs appears no different between the two, however, with seven or eight PCs along both stretches of the cold front. These differences are possibly related to the stronger cold front and hence greater along-frontal vertical shear at the poleward than equatorward end (compare Figs. 6b,d). There is also less static stability above the cold front at the equatorward than poleward end (Figs. 6b,d), which may be a factor in the more convective appearance of the PCs at the equatorward end.

Twelve hours later, the precipitation at the poleward end has fully become a smooth, continuous cold-frontal rainband (Fig. 7a). Although, as shown in section 3a, the PCs become less pronounced over time at the equatorward end, they are still readily evident in the equatorward box (Fig. 7c). The same contrasts in near-surface along-front vertical shear and static stability (compare Figs. 7b,d) between the poleward and equatorward boxes as previously are evident. Thus, the PCs are more pronounced and persist longer at the equatorward than poleward end.

4. Effect of latent-heat release on precipitation cores

This section shows the effect on the PCs of removing latent-heat release from the microphysics, by comparison of CNTL and NOLATENT, which are equivalent except for the absence of latent-heat release in NOLATENT. After 18 h in the equatorward box, the PCs that form distinctively in CNTL, as well as the cold-frontal rainband itself, have not formed in NOLATENT (Figs. 8a,c). At this time, the wind shift across the cold front in NOLATENT is less sharp than in CNTL (compare Figs. 8a,c). The cold-frontal magnitude and along-frontal vertical shear are also notably weaker (compare Figs. 8b,d). Furthermore, the static stability in NOLATENT is greater than in CNTL (compare Figs. 8b,d), which may be a factor in the lower precipitation rate along the cold front.

Twelve hours later, however, PCs have formed in NOLATENT (Fig. 9c), and there is less of a contrast in θ-gradient and vertical shear between the two simulations (compare Figs. 9b,d) than previously. Although the PCs in CNTL are not at much greater an angle to the cold front than in NOLATENT, the wavelength of the PCs is much greater in CNTL, with 6 or 7 PCs in CNTL, compared to 12 or 13 in NOLATENT (Figs. 9a,c). Thus, the PCs in NOLATENT are much less distinctive and more closely resemble a continuous cold-frontal rainband.

This contrast between CNTL and NOLATENT is similar to that between the poleward and equatorward ends of the cold front, shown in section 3c. However, in that case, the contrast was due to a greater angle of the PCs at the equatorward than poleward end, with the wavelength roughly the same between the two ends. In this instance, the PCs are at no greater an angle in CNTL than NOLATENT, but the wavelength is much greater. As in Kawashima (2011), this increased wavelength between simulations is associated with greater wind shift across the front.

5. Effect of surface friction on precipitation cores

In this section, a land mass is created, occupying the righthand side of the outer domain and most of the nested domain (Fig. 2). CNTL, in which the land mass does not exist and the cold front continues to move over the ocean, is compared to LANDFRIC1, LANDFRIC2, and LANDFRIC3, in which the roughness length of the land mass is increased from LANDFRIC1 to LANDFRIC3. Thus, this section illustrates the sensitivity of PCs to surface friction.

Figure 10 shows the time series of domain-maximum surface relative vorticity and near-surface along-frontal vertical shear in these four simulations. The moment that the cold front crosses the coastline in LANDFRIC1, LANDFRIC2, and LANDFRIC3 is readily evident in both time series, with the vorticity (Fig. 10a) and vertical shear (Fig. 10b) both dropping, having increased while approaching the coastline, which is more pronounced with greater friction. The only exception is that the vertical shear in LANDFRIC3 remains greater after crossing the coastline than in LANDFRIC2.

The vorticity reduction over land is further illustrated in Fig. 11, in which the surface winds and hence the magnitude of the wind shift across the cold front are weaker when friction is greater. In simulations with greater friction, the PCs are longer and at a greater angle to the cold front, in contrast to the effect of latent heat, which only affected the length of PCs. Thus, in LANDFRIC2 and LANDFRIC3, there are much larger gaps between PCs than in any previous simulations.

Figure 12 shows differences between these four simulations, which appear to be strongly influencing the morphology of the PCs. The surface cold front is weaker in simulations with greater friction, so that the near-surface vertical shear is weaker, and increasingly oriented along and increasingly oriented across the cold front. The verti-
6. Effect of sea-surface temperature on precipitation cores

In this section, the coastline that was used in section 5 to vary surface friction between simulations is used to vary the SST distribution of the lower boundary that the cold front moves over, keeping roughness length the same between simulations. Thus, this section investigates the effect on PCs of higher SST and, thus, greater surface sensible- and latent-heat fluxes.

The effect of crossing the coastline on surface vorticity is less for these simulations than when the coastline contains a frictional discontinuity (Fig. 10a), but there is a slight increase in vorticity between about 15 h and 30 h as SST increases between simulations (Fig. 14a). More striking in the SST-sensitivity simulations is the effect on vertical shear. There is no clear signal in the time series of maximum $\frac{\partial u}{\partial z}$ (not shown), but the time series of maximum $\frac{\partial v}{\partial z}$ shows a large reduction in simulations with higher SST between about 10 h and 30 h. Therefore, in these simulations, SST has no major impact on the along-front vertical shear, but higher SST greatly reduces the across-front vertical shear.

After 18 h, there is little difference between the five simulations (Fig. 15). However, in MINUS5K, the PCs along part of the cold front are longer and overlap more than in the other simulations, while, in PLUS5K, along the same part of the cold front, the PCs are poorly defined and show signs of forming a continuous rainband.

Twelve hours later, the influence of SST is much clearer. The simulations have increasingly intense precipitation along
Fig. 11. Comparison of equatorward boxes in CNTL (at 18 h), LANDFRIC1 (at 20 h), LANDFRIC2 (at 22 h), and LANDFRIC3 (at 24 h). Panels are as lefthand panels of Figs. 4, 6, 7, 8, and 9. Simulations are presented at different times to one another to allow for slower evolution of PCs with greater friction.
Low. trop. lapse rate (greys), sfc. \( \theta \) (red), 950-hPa vert. shear (vectors).

Fig. 12. As Fig. 11, but panels are as righthand panels in Figs. 6, 7, 8, and 9.
Sfc. precip. (colored), $\theta$ (grey), winds.

Fig. 13. As Fig. 11, but 12 h later in each simulation (CNTL at 30 h, LANDFRIC1 at 32 h, LANDFRIC2 at 34 h, and LANDFRIC3 at 36 h).
the cold front with higher SST (Fig. 16), in accordance with greater lapse rate above and ahead of the cold front (Fig. 17), as is to be expected for greater SST. Unlike the previous sensitivity simulations, higher SST has also enhanced the width of PCs (Fig. 16). However, higher SST has caused the PCs to be more along-front oriented, so that, in PLUS2K and PLUS5K, the PCs have reverted to a single continuous rainband. This effect appears to be related to the contrast in vertical shear between the simulations, illustrated in Fig. 14; Figure 17 shows that the large shear normal to and just ahead of the cold front in MINUS5K is gradually smaller between simulations, so that, in PLUS5K, the only large shear is along the front. Therefore, as in the friction-sensitivity simulations, large cross-frontal vertical shear, relative to along-frontal vertical shear, is associated with high-amplitude PCs.

Thus, in these sensitivity simulations, differences in SST at the coastline take longer to take effect than in the surface-friction sensitivity simulations in section 5. However, prescribing higher SST eventually impedes the PCs and, despite enhancing their width and intensity, there is a tendency for PCs to form a continuous rainband. In this way, greater SST has the opposite effect to latent heat (section 4) and friction (section 5), both of which enhance PCs, leading to a less uniform cold-frontal rainband.

7. Summary

Moist idealized baroclinic-wave simulations were performed, in which a mature cold front of maritime origin was re-initialized with a nested domain of 4-km grid spacing inserted to simulate clockwise-oriented precipitation cores (PCs) along the cold front. PCs form along the full length of the cold front, in conjunction with counter-propagating wave-like vorticity maxima either side of the cold front, possibly due to horizontal shear instability. However, the PCs remain near steady-state over about 18 hours, gradually becoming less pronounced, more closely resembling a continuous narrow cold-frontal rainband. Thus, the PCs in a more realistic model are more complex than horizontal-shear-instability theory can explain.

Nevertheless, the PCs appear to have similar sensitivities to horizontal shear, vertical shear, and static stability as precipitation structures, resembling observed PCs, simulated from horizontal shear-instability by Kawashima (2011). As in his simulations, generally:

- greater cross-frontal horizontal shear (wind shift across the cold front) is associated with greater wavelength of the PCs;
- greater cross-frontal, relative to along-frontal, vertical shear is associated with greater angle of the PCs to the cold front;
- greater lapse rate above the cold front is associated with more poorly organized PCs.

The only exception to the above was that greater friction between simulations led to smaller wind shift, but greater wavelength. However, Kawashima also simulated PCs with greater wavelength by increasing the cross-frontal vertical shear between simulations, and this relationship was also evident in our friction-sensitivity simulations. Therefore, in the friction-sensitivity simulations, the vertical shear appears to have controlled the PCs.
Sfc. precip. (colored), $\theta$ (grey), winds.

Fig. 15. As Fig. 11, but comparing simulations of different SST. Unlike Fig. 11, simulations are all presented at the same time (18 h).
Fig. 16. As Fig. 15, but after 27 h (9 h later).
Low. trop. lapse rate (greys), sfc. θ (red), 950-hPa vert. shear (vectors).

Fig. 17. As Fig. 16, also after 27 h, but panels are as Fig. 12.
However, the primary purposes of this paper are to showcase the PCs that can be simulated along a cold front in a more realistic idealized model than has been used previously to investigate PCs, and to illustrate their sensitivity to latent-heat release, surface friction, and sea-surface temperature (SST). The PCs are most distinctive at the equatorward end of the cold front, lying at a greater angle to the cold front than at the poleward end, so that the gaps along the cold front are larger. The well-defined PCs at the equatorward end persist longer than the poorly defined PCs at the poleward end.

Latent heat and surface friction enhance PCs, while higher SST inhibits them. However, the manner in which each of these factors affects PCs is different. Latent-heat release does not greatly increase the angle of PCs to the cold front, but makes the PCs much longer, with larger gaps along the cold front. Greater surface friction increases the length, angle, and curvature of PCs. Therefore, when friction is large, there are very long gaps of lighter or no rainfall along the cold front, and the most distinctive PCs form, such as in Fig. 1a. Despite enhancing the width and intensity of the cold-frontal rainband, higher SST reduces the angle of PCs to the cold front and makes the rainband more uniform. These sensitivity simulations may explain much of the observed variability in PCs, both along the cold front within an individual case and between different cases.

Acknowledgements

The precipitation-radar data plotted in Fig. 1 were provided by the Met Office through the British Atmospheric Data Centre (BADC). Thanks are given to Riwal Plougonven, who provided the code to produce the initial jet, and Tim Slater, who altered the code to produce a jet at 20-km grid spacing. Jesse Norris is a NERC-funded student through the DIAMET (DIAbatic in ExTratropical storms) project, NE/I005234/1.

REFERENCES


Paper 3: Snowbands over the English Channel and Irish Sea during cold-air outbreaks.
Persistent northerly-to-easterly cold-air outbreaks affected the UK during the winters of 2009–10 and 2010–11, with the resulting convection frequently organizing into snowbands over the English Channel and Irish Sea. Sounding data and composite radar reflectivity images from the Met Office Nimrod precipitation radar network reveal that these bands formed along the major axis of each body of water (or sea) when the boundary-layer flow was roughly parallel to each of those axes (along-channel). For both seas, a band was present the majority of times that the 850 hPa flow was along-channel. Of these times of along-channel flow, the 850 hPa wind speed and surface-to-850 hPa temperature difference were significantly greater when bands were present than when they were not. For the English Channel only, the land–sea temperature difference was also significantly greater when bands were present than when they were not. In a real-data Weather Research and Forecasting model (WRF) control simulation of a typical band over the English Channel, a trough develops over the water and offshore air streams from either side converge along it. In the absence of surface fluxes, the trough, convergence and organized precipitation fail to develop altogether. Orography and roughness-length variations are less important in band development, affecting only the location and morphology.

Key Words: lake effect; land breeze; radar; WRF; convection; snowbands; cold-air outbreaks

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1. Introduction

During the winters of 2009–2010 and 2010–2011, anti-cyclonic blocking over the North Atlantic led to an anomalous synoptic-scale flow regime over northern Europe, with extremely cold, dry air being advected over the UK from the north and east. With the polar jet typically a long way south, often as far as the Mediterranean Sea, fewer cyclones affected the UK than during a westerly regime and hence synoptic-scale fronts were less of a factor in generating precipitation. Instead, the precipitation during these periods was characterized by clusters of convective cells arriving from the North Sea and organizing around the UK’s coasts (Figure 1). Snow resulting from this convection fell in the same locations for several consecutive days. With temperatures barely rising above freezing during the daytime, snow accumulated as Arctic air flowed over the UK. In January 2010 and December 2010 (the culmination of the blocking in each of the winters), almost the whole country experienced at least eight and twelve days of snow cover, respectively (http://www.metoffice.gov.uk/climate/uk/anomacts/). In both winters, snow depths exceeding 10 cm were widespread around the country with depths up to 50 cm in some locations over higher ground (http://www.metoffice.gov.uk/climate/uk/interesting/). Leading insurance company RSA estimated that the severe winter weather in late 2010 cost the UK economy £1.2 billion per day (http://www.channel4.com/news/snow-chaos-costs-uk-economy-1-2bn-a-day). In the final quarter of 2010, GDP fell by 0.5%, which the Office for National Statistics attributed to the weather

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Although most of the convection during the cold-air outbreaks in these winters occurred over the eastern UK, distinctive precipitation features formed over the English Channel and Irish Sea (Figure 2), where convective cells frequently organized into bands. Snowbands were elongated parallel to the major axis of the relevant body of water (hereafter ‘sea’), spanned the majority of the sea’s length and remained quasi-stationary, often for longer than 24 h. They formed over each sea when the boundary-layer flow was parallel to its major axis (approximately eastnortheasterly for the English Channel and approximately northnortheasterly for the Irish Sea), hereafter ‘along-channel’. Although they formed over water, parts of these bands regularly came onshore (Figures 3 and 4), leading to snowfall in the same locations for many hours. They were either ‘single bands’, where precipitation organized along a single axis (Figures 3(a) and 4(a)) or ‘multibands’, where precipitation organized along two or more parallel axes (Figures 3(b) and 4(b)). One particular place that was hard hit was Guernsey, an island in the English Channel, where thundersnow occurred (Guernsey Meteorological Office, 2011) as the cold air over warm waters produced the necessary conditions for electrification and snow (e.g. Schultz, 1999; Schultz and Vavrek, 2010).

That such bands form during cold-air outbreaks over water suggests that the heat and moisture fluxes from the water are crucial to their formation. Indeed, shallow convection is generated over warm water when cold air in the boundary layer flows over it (Benard, 1900; Rayleigh, 1916). Over an open ocean, the resulting convection typically takes one of two morphologies—cellular convection or horizontal convective rolls—depending upon the static stability and vertical shear of the horizontal wind (hereafter vertical wind...
obtained a widely varying precipitation distribution when
plays a role. To that effect, Andersson and Gustafsson (1994)
garner different results on the extent to which each of these
However, modelling studies over different bodies of water
land–water contrasts have also been studied along the New
addition to a shore on either side of the flow) and a sea does
here is that a lake has an upwind and downwind shore (in
difference between bands over lakes and the ones studied
et al.
bands (Laird et al., 2003a,b).
Shoreline bands forming along the major axis of a lake are
relevant to this study; over the Great Lakes (e.g. Peace and
Sykes, 1966; Passarelli and Braham, 1981; Hjelmfelt, 1990) and over the Great Salt Lake (Steinburch et al., 2000; Steinburgh and Onton, 2001; Onton and Steinburgh, 2001). The single bands in Figures 3(a) and 4(a) (and other single bands identified in this study) are similar to the Type IV lake-effect snowband of Nizioł et al. (1995). The crucial difference between bands over lakes and the ones studied here is that a lake has an upwind and downwind shore (in addition to a shore on either side of the flow) and a sea does not. However, wintertime wind-parallel bands arising from land–water contrasts have also been studied along the New England coast (Bosart, 1975), off the north coast of Germany (Pike, 1990), over the Sea of Japan (Nagata, 1987) and over the Baltic Sea (e.g. Andersson and Nilsson, 1990; Andersson and Gustafsson, 1994). Some combination of the following factors is generally concluded to have been responsible for generating lift for the observed banding:
  - thermally driven land breezes;
  - frictional differences between land and sea;
  - deflection of air around orography.

However, modelling studies over different bodies of water garner different results on the extent to which each of these plays a role. To that effect, Andersson and Gustafsson (1994) obtained a widely varying precipitation distribution when
altering the geometry of the coastline around the Baltic Sea, coining the terms 'coast of departure' and 'coast of arrival'. Thus, the unique geography of each body of water implies that each warrants its own investigation into how bands form there.

The precipitation band over the Irish Sea has been
given a name by weather enthusiasts: the Pembroke
or Pembrokeshire Dangler. There is some discrepancy
over how the band forms, however. For example, one
source attributes the band to convergence produced by deflection of flow around the Pembrokeshire peninsula (Figure 2: http://weatherfaqs.org.uk/node/216), whereas another attributes the band to northerly flow through the North Channel meeting converging land breezes, forcing convergence the length of the Irish Sea (http://en.wikipedia.org/wiki/Pembrokeshire Dangler). Given the lack of agreement over the origins of these bands, how they form would seem to be a topic ripe for investigation.

To date, there has not been a comprehensive description of such bands around the UK in the open scientific literature. The nearest candidate is the investigation of Browning et al. (1985), who identified wind-parallel cloud bands over the North Channel (Figure 2) and Irish Sea. This study was followed up by Monk (1987) and Monk et al. (1990); their results were not published in a scientific journal but images from Monk (1987) appeared in Bader et al. (1995, Section 6.2.4). These studies first called attention to these features and Monk (1987) identified the three lifting mechanisms itemized above as possible causes.

As far as we know, the scientific literature contains neither a climatology of atmospheric conditions associated with these UK bands nor a modelling sensitivity study of them. Thus, the extent to which the mechanisms forming the snowbands resemble those of lake-effect bands or those referred to over other bodies of water around the world remains unknown. This article aims to address these issues. The availability of precipitation radar observations of these bands and high-resolution mesoscale models to simulate them allows us to study these bands in a way not done to date.

The remainder of this article is organized as follows. Section 2 outlines the data and methods. Section 3 gives a case study of a typical band over each sea during the winters in question. In section 4, synoptic composites are presented for the times when bands initiated. In section 5, a climatology of sounding data is presented. In section 6, real-data simulations are presented for a band over the English Channel. In section 7 we discuss the results and compare them with previous literature. In section 8 we summarize the findings.

2. Data and method

Met Office precipitation radar (Nimrod) composites were examined for every day during periods of 'snow and low temperatures', as defined on the Met Office's website (http://www.metoffice.gov.uk/climate/uk/interesting/). These were from 17 December 2009–15 January 2010, 25 November 2010–9 December 2010 and 16–26 December 2010. Radar data were unavailable from 20 December 2010 onwards due to a data-processing error at the Met Office. Therefore, 49 days were available for analysis.

Figure 4. As Figure 3 but over the Irish Sea: (a) a single band at 2145 UTC on 18 December 2009; (b) a multiband at 0815 UTC on 25 November 2010.
Times were recorded at which bands initiated along the major axis of each sea, excluding those that formed roughly along a synoptic-scale front. This approach identified 14 bands over the English Channel and 19 over the Irish Sea. In each case, the low-level winds were broadly along-channel; synoptic composites of these initiation times are presented in section 4.

To establish why bands sometimes did not form, given favourable wind direction, the study identified the times during these 49 days when the 850 hPa flow was along-channel. These times were diagnosed from sounding data near the wind end of the English Channel and Irish Sea, respectively (Figure 2). For the English Channel, a sounding from Herstmonceux and Castor Bay, which are located near the upwind end of the English Channel and Irish Sea, respectively (Figure 2). For the English Channel, a sounding from Herstmonceux was included in the study if the 850 hPa wind direction was between 45° and 90° (i.e. between northeasterly and easterly), which we hereafter term eastnortheasterly or ENE flow. For the Irish Sea, the requirement was that the 850 hPa wind direction over Castor Bay be between 0° and 45° (i.e. between northerly and northeasterly), which we hereafter term northnortheasterly or NNE flow.

These times were at which bands initiated along the English Channel and 16 times of NNE flow over the Irish Sea. For each of these sounding times, the corresponding radar precipitation image was studied to establish whether a band was present or not and, if so, whether it was single or multiband.

### 2.1. Nimrod data

Nimrod composite maps are constructed from the radar reflectivity measurements from 18 5.6 cm wavelength radars, evenly distributed around the British Isles (http://www.metoffice.gov.uk/weather/uk/radar/tech.html). The data from each radar are sent to the Met Office headquarters and converted into a 1 km × 1 km grid of reflectivity values (Z, in mm$^6$ m$^{-3}$), which are then converted to precipitation rate (R, in mm h$^{-1}$) using the equation in Table 1 of Harrison et al. (2009):

$$Z = 200R^{1.6}. \quad (1)$$

An example is plotted in Figure 1.

Radar data, however, have their limitations. Problems arise at long ranges because, due to the curvature of the Earth, each radar can only observe precipitation some distance above the surface. For example, precipitation may evaporate below the lowest elevation scan, leading to an overestimate of precipitation at the surface. Also, the radar beam may pass over the top of low-lying clouds, underestimating precipitation. As shown in section 3, this underestimate may be occurring with these bands because of the shallow convection.

Another common error is the bright band, in which the high reflectivity of droplets detected at the level where snow is melting returns strong echoes, leading to an overestimate of intensity. However, the vast majority of the radar echoes observed in this study were from precipitation that reached the surface as snow and so this is unlikely to have occurred.

In this study, we are primarily concerned with the morphology, as opposed to intensity, of precipitation. Bands were manually identified from criteria for orientation, length, width and intensity in Table 1. Additional criteria were applied to determine multibands (Table 2). Multibands usually consisted of one longer band accompanied by one or more shorter bands. Thus, to qualify as a multiband, the threshold for length was less for any additional parallel bands. Any band satisfying the criteria in Table 1 but not Table 2 was considered a single band.

These criteria led to 10 times at which a band was observed over the English Channel (from the 19 times of ENE flow at 850 hPa) and 10 times at which a band was observed over the Irish Sea (of the 16 times of NNE flow at 850 hPa). Of these, 8 over the English Channel and 4 over the Irish Sea were multibands. A climatology of sounding data, using these times of along-channel flow, is presented in section 5.

### 2.2. Sea-surface temperatures

Sea-surface temperatures (SSTs), taken from Met Office buoys and light vessels over the two seas, were also used to inform the climatology. At each time of ENE and NNE flow, the mean SST was calculated. The surface temperature taken from the sounding was subtracted from this mean SST to obtain the local land–sea temperature difference, $\Delta T$, and thus assess the impact of thermally driven circulations on the generation of bands. Similarly, the 850 hPa temperature taken from the sounding was subtracted from the mean SST to obtain the surface-to-850 hPa temperature difference and thus assess boundary-layer stability over the water. Following Braham (1983), a surface-to-850 hPa temperature difference of about 13 K is sufficient for free convection as this corresponds to the dry adiabatic lapse rate. However, as shown in section 3, some cloud bases for these bands were

### Table 1. Threshold criteria for single bands identified from Nimrod images.

<table>
<thead>
<tr>
<th>Parameter</th>
<th>English Channel</th>
<th>Irish Sea</th>
</tr>
</thead>
<tbody>
<tr>
<td>Orientation</td>
<td>A line or curve of approximately eastnortheast–westsouthwest orientation</td>
<td>A line or curve of approximately northnortheast–southsouthwest orientation</td>
</tr>
<tr>
<td>Length</td>
<td>At least half the distance between the Dover–Calais and Penzance–Brest midpoints (Figure 2)</td>
<td>At least half the distance between Douglas and the Penzance–Cork midpoints (Figure 2)</td>
</tr>
<tr>
<td>Width</td>
<td>$\leq 50$ km at all points along length</td>
<td></td>
</tr>
<tr>
<td>Intensity</td>
<td>$\geq 1$ mm h$^{-1}$ at regular intervals along identified line or curve</td>
<td></td>
</tr>
</tbody>
</table>

### Table 2. Threshold criteria for multibands identified from Nimrod images.

<table>
<thead>
<tr>
<th>Parameter</th>
<th>English Channel</th>
<th>Irish Sea</th>
</tr>
</thead>
<tbody>
<tr>
<td>Orientation</td>
<td>As in Table 1 for at least two parallel bands</td>
<td>As in Table 1 for longest band and at least half of that distance (a quarter of the length of the relevant sea) for at least one other</td>
</tr>
<tr>
<td>Length</td>
<td>As in Table 1 for at least two parallel bands</td>
<td></td>
</tr>
<tr>
<td>Width</td>
<td>As in Table 1 for at least two parallel bands</td>
<td></td>
</tr>
<tr>
<td>Intensity</td>
<td>As in Table 1 for at least two parallel bands</td>
<td></td>
</tr>
<tr>
<td>Spacing</td>
<td>$\geq 15$ km between parallel bands where intensity is continuously, or near-continuously, $\leq 1$ mm h$^{-1}$</td>
<td></td>
</tr>
</tbody>
</table>
below 850 hPa so the associated convection in some cases could have resulted from smaller temperature differences.

3. Case studies

To illustrate typical band development and decay, two case studies are presented: one over the English Channel (section 3.1) and one over the Irish Sea (section 3.2).

3.1. English Channel

A band developed and decayed over the English Channel on 30 November and 1 December 2010 (Figure 5). A zonally elongated high over Scandinavia and a zonally elongated low over western Europe led to parallel, near-zonally oriented isobars across the UK (Figure 6). The Met Office analysis also features multiple troughs, predominantly over water, the longest of which is along the English Channel (Figure 6). The Herstmonceux sounding at 0000 UTC on 1 December indicates northeasterly flow below 500 hPa (Figure 7). Temperature and dew-point temperatures in the boundary layer were below 0°C with a capping inversion at 750 hPa overlying a shallow cloud deck about 100 hPa thick (Figure 7).

Radar echoes began to organize along the English Channel about 0800 UTC on 30 November. At 1200 UTC, as the band was beginning to form (Figure 5(a) and (e)), the 850 hPa wind speed was 11.8 m s\(^{-1}\), up from 8.2 m s\(^{-1}\) at 0000 UTC, and eastnortheasterly (Table 3). Initially, the echoes exhibited a multibanded structure of eastnortheast–westsouthwest orientation, occupying only the western half of the English Channel (Figure 5(a)). Visible

\[ T - \log p \] chart at Herstmonceux at 0000 UTC on 1 December 2010. Courtesy of the University of Wyoming. This figure is available in colour online at wileyonlinelibrary.com/journal/qj
Table 3. Evolution of variables as diagnosed from SST data over the English Channel and sounding data at Herstmonceux during the development and decay of a band over the English Channel.

<table>
<thead>
<tr>
<th>Time &amp; date</th>
<th>‘Instability’</th>
<th>(U_{850})</th>
<th>(\Delta T)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Pre-initiation 00 UTC 30 Nov</td>
<td>19.2</td>
<td>8.2 (60°)</td>
<td>11.5</td>
</tr>
<tr>
<td>Developing 12 UTC 30 Nov</td>
<td>20.9</td>
<td>11.8 (80°)</td>
<td>11.9</td>
</tr>
<tr>
<td>00 UTC 1 Dec</td>
<td>22.1</td>
<td>16.4 (60°)</td>
<td>12.7</td>
</tr>
<tr>
<td>Decaying 12 UTC 1 Dec</td>
<td>22.2</td>
<td>18.0 (40°)</td>
<td>13.5</td>
</tr>
</tbody>
</table>

Variables are surface-to-850 hPa temperature difference (‘instability’), 850 hPa wind (\(U_{850}\)) and land–sea temperature difference (\(\Delta T\)). Whether the band had not yet initiated, was developing, decaying or no longer visible is indicated at the time of each sounding.

satellite imagery confirms the presence of separate parallel cloud streets at this time (Figure 5(e)).

Between 1200 UTC on 30 November and 0000 UTC on 1 December, eastnortheasterly flow strengthened, reaching 16.4 m s\(^{-1}\) at 850 hPa (Table 3). Although the intensity of the radar echoes did not increase, the band became better defined and lengthened, occupying the full length of the English Channel by 0200 UTC (Figure 5(b)). The separate parallel bands became less distinctive and gradually merged into one wider band (Figure 5(b)). The infrared imagery at this time shows distinctive cloud along part of the English Channel (Figure 5(f)), but it is not until the visible imagery becomes available at 0800 UTC that the reduction to a single band is seen on satellite (Figure 5(g)), at which time the radar shows the band still well defined and spanning the full length of the English Channel (Figure 5(c)). At this time, the location of the band corresponds exactly to the trough on the surface pressure analysis (Figure 6), the latter likely drawn by a Met Office forecaster observing the location of the cloud and radar echoes.

Between 0000 UTC and 1200 UTC on 1 December, 850 hPa wind speed further increased to 18 m s\(^{-1}\) but became more northerly, no longer quite parallel to the English Channel (Table 3). During this time, the band steadily dissipated (observe the transition from Figure 5(b) to Figure 5(d)). The band gradually rotated anti-clockwise throughout its evolution (observe the transition from Figure 5(a) to Figure 5(d)), roughly following the direction of the wind. During its decay, the band regained its multibanded structure (Figure 5(d) and (h)).

Throughout the band’s development and decay, the surface-to-850 hPa temperature difference was between 20.9 and 22.2 K (15.7–17.6 °C km\(^{-1}\), thus absolutely unstable) and the land–sea temperature difference was between 11.9 and 13.5 K (Table 3). However, there was no marked change in these variables during development or decay (they both increased very gradually during both development and decay).

Figure 8. Nimrod data at 1 km grid spacing ((a)–(d), mm h\(^{-1}\)), and visible ((f), (g)) and infrared ((e), (h)) satellite imagery from Meteosat Second Generation, charting the development and decay of a band over the Irish Sea. (a) and (e): 2300 UTC on 6 January 2010; (b) and (f): 0900 UTC on 7 January 2010; (c) and (g): 1600 UTC on 7 January 2010; (d) and (h): 2200 UTC on 7 January 2010. Domain as in Figure 4.
3.2. Irish Sea

A band developed and decayed over the Irish Sea on 6–7 January 2010 (Figure 8). High pressure over the North Atlantic induced northerly flow along the Irish Sea and again troughs were analyzed, including over the Irish Sea (Figure 9). Near-surface dewpoints were around \(-5^\circ\)C at Castor Bay, implying little moisture, and in this case the cloud top was only at about 800 hPa (Figure 10), suggesting shallower convection than in the English Channel case.

Radar echoes began to organize along the Irish Sea about 2300 UTC on 6 January (Figure 8(a)). At 0000 UTC on 7 January, the 850 hPa wind was 9.8 m s\(^{-1}\) and northerly (Table 4). Initially, a multiband lay just south of Douglas (Figure 2), its two constituent bands of different orientation from one another (Figure 8(a) and (e)).

Thereafter, the multibands merged and, until about 0900 UTC, steadily lengthened but did not widen or intensify, by which time the single band spanned almost the full length of the Irish Sea (Figure 8(b) and (f)). Sounding data are unavailable until 1200 UTC, at which time the 850 hPa wind was 15\(^\circ\), still parallel to the Irish Sea, but its speed had plummeted to 4.1 m s\(^{-1}\) (Table 4).

Indeed, at this time and thereafter, the band was steadily decaying (Figure 8(c) and (g)). Although not detected on radar (Figure 8(c)), the visible imagery at 1600 UTC shows that the band had separated into two parallel cloud streets (Figure 8(g)).

At 0000 UTC on 8 January, the 850 hPa wind was 3.6 m s\(^{-1}\) (Table 4).

The surface-to-850 hPa temperature difference was between 16.3 and 18.1 K (11.2–12.9 °C km\(^{-1}\), thus absolutely unstable) and the land–sea temperature difference was between 10.9 and 11.6 K throughout the band’s development and decay (Table 4). As in the English Channel case, there was no great variability in these variables (surface-to-850 hPa temperature difference, in fact, gradually decreased throughout).

<table>
<thead>
<tr>
<th>Time &amp; date</th>
<th>‘Instability’</th>
<th>(U_{850}) (m s(^{-1}))</th>
<th>(\Delta T) (K)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Developing</td>
<td>00 UTC 7 Jan</td>
<td>18.1</td>
<td>9.8 (350(^\circ))</td>
</tr>
<tr>
<td>Decaying</td>
<td>12 UTC 7 Jan</td>
<td>17.2</td>
<td>4.1 (15(^\circ))</td>
</tr>
<tr>
<td></td>
<td>00 UTC 8 Jan</td>
<td>16.3</td>
<td>3.6 (25(^\circ))</td>
</tr>
</tbody>
</table>

Variables and annotations are as in Table 3.

3.3. Synthesis

Both bands formed in strong along-channel winds (about 12 and 10 m s\(^{-1}\) at 850 hPa for the English Channel and Irish Sea, respectively). In the English Channel case, the 850 hPa wind speed continued to increase during the decay but its direction changed, no longer along-channel. In the Irish Sea case, the 850 hPa wind remained along-channel but slowed to about 4 m s\(^{-1}\). In both cases, the surface-to-850 hPa temperature difference and land–sea temperature difference were high, implying an unstable boundary layer and large differential heating between land and water. However, there was no marked increase in these variables during development or decrease during decay. Thus, in both cases, the evolution of the wind field appears to have controlled the development and decay of the bands.

The two cases were also similar in that they both initiated as poorly defined multibands. Over about 12 h, the multibands merged to form a single smooth, almost perfectly straight band, spanning the length of the relevant sea. Subsequently, the single band became increasingly patchy and re-formed multibands, dissipating over about 12 h. Many other single bands over both seas during the winters in question started and ended as multibands. However, some single bands did not evolve from or into multibands and some multibands did not evolve into single bands.

4. Synoptic composite analysis

To illustrate the synoptic-scale weather patterns associated with band formation, synoptic composites are created from reanalysis data from the National Centers for Environmental Prediction (NCEP) and the National Center for Atmospheric Research (NCAR) (Kalnay et al., 1996) using the compositing web site created by the

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5. Climatology

Section 4 illustrates the synoptic-scale environment in which the observed bands formed. However, as stated, bands failed to form at other times in a similar synoptic-scale environment. Indeed, for both seas, banding was present only a slight majority of times that the 850 hPa wind direction was along-channel (Table 5). We thus turn our attention to differences between times at which bands were present and times at which they were not.

Over both seas, the mean surface-to-850 hPa temperature difference was greater at times of banding than at times of no banding (Table 5). This difference is greater over the English Channel than over the Irish Sea (3.4 K compared with 1.3 K). However, in both the English Channel and Irish Sea cases this is significant at the 95% level, using the t-test (e.g. Wilks, 1995, pp122–124; hereafter just ‘t-test’).

Over both seas, the mean 850 hPa wind speed was more than 4 m s$^{-1}$ greater at times of banding than at times of no banding (Table 5). For both seas, this difference is significant at the 95% level using the t-test.

For the English Channel, $\Delta T$ was just 1.4 K greater at times of banding than at times of no banding (Table 5), but this is significant at the 95% level using the t-test. However, for the Irish Sea, $\Delta T$ was 4.4 K greater at times of no banding than at times of banding, which is significant at the 95% level using the t-test. We are unable to offer an explanation for this unexpected result.

The variables in Table 5 were also compared between times at which single and multibands were observed. No significant results were obtained to distinguish between the two, however.

6. Real-data simulations

According to the ingredients-based approach, moist convection requires the simultaneous presence of instability, moisture and lift (e.g. Johns and Doswell, 1992). Table 5 demonstrates that the lower atmosphere was sufficiently unstable for the formation of these bands and, despite the relative dryness of the cold-air outbreaks, their formation over relatively warm water explains the origin of the moisture. Thus, the lifting mechanisms responsible for the observed bands are now investigated.

From Table 5, bands were favoured by faster winds and, for the English Channel, by greater land–sea temperature contrast. How did these factors lead to convergence and lift? To examine the extent to which each of the three lifting mechanisms itemized in section 1 was responsible for the observed bands, model simulations of one such band were conducted (that which occurred over the English Channel between 30 November and 1 December 2010, presented in section 3.1).

6.1. Model setup

Simulations were performed with version 3.3.1 of the Weather Research and Forecasting (WRF) model with...
the Advanced Research WRF (ARW) dynamical core (Skamarock et al., 2005). Three model domains were set up using Lambert conformal mapping (Figure 12) with two-way nesting, grid spacings of 25, 5 and 1 km and time steps of 150, 30 and 6 s. 58 model eta levels were used. European Centre for Medium-Range Weather Forecasts (ECMWF) analysis data (ECMWF, 1994) at 0.5° latitude–longitude grid spacing were used for input into the model every 6 h. NOAA SST analysis data (Reynolds, 1988) at 0.5° latitude–longitude grid spacing, available every 24 h, were interpolated for input into the model at the same times.
Simulations were started at 1200 UTC on 28 November 2010, about 48 h before the band initiated. The finish time was 0000 UTC on 2 December, about three hours after the band dissipated. Grid-scale noise was removed by adding sixth-order numerical diffusion.

A full-physics control run was performed, followed by a run in which surface heat and moisture fluxes were switched off to establish the importance of differential heating between land and sea. Then, simulations were performed in which orography, land–sea frictional differences and surface fluxes were removed cumulatively. In the case of land–sea frictional differences, a constant value of roughness length, $z_0 = 0.2$ mm (that of open sea), was set. Thus, five simulations were performed (Table 6), the results of each of which are described in turn.

### 6.2. Control run

A sea-level-pressure trough forms over the English Channel (Figure 13(b)). At 0600 UTC on 30 November (42 h into the simulation), as the radar shows the band initiating (Figure 13(a)), winds turn towards the trough in the model simulation as air flows over southern England (Figure 13(b)). Winds across the north coast of France also flow into the trough, meeting the air from southern England. A mesoscale vortex thus lies over the trough, slightly to the west of where it appears to be from radar (Figure 13(a)). Multiple convergence zones (Figure 13(b)) and reflectivity bands (Figure 14(b)) surround the vortex, arising from the confluence of the two air streams. Greater wind speed over water than land further intensifies the convergence. These bands of convergence correspond to bands of divergence higher up, with maxima at about 850 hPa and fading away at about 550 hPa (not shown). Land–surface-station observations also capture the turning of winds over England and convergence over the English Channel (Figure 13(a)).

Subsequently the trough becomes more uniform and 12 h later, at 1800 UTC, the orientation of the isobars is much the same along the length of the English Channel (Figure 15(b)). The 10 m wind field evolves into a uniform confluence at the downwind end of the English Channel of the two air streams from the north and south, which is also exhibited by the land-surface-station wind vectors (Figure 15(a)). Thus, in the model, the reflectivity is organized into a smooth single band along the downwind end of the English Channel (Figure 15(b)), similar to that on radar (Figure 15(a)), although this single-band transition occurs slightly earlier than in the radar imagery (the morphology is more distinctively multibanded in the radar imagery than in the model at this time).

### 6.3. No surface fluxes

In the absence of surface fluxes, the trough does not develop at 0600 UTC (Figure 14(c)) or 1800 UTC (Figure 15(c)). There is still a turning of winds towards low pressure as they cross the east coast of England (Figure 13(c)) and the airflow offshore from southern England is not much different visibly from that in the control run. However, there is no offshore flow across the north coast of France and the confluence observed in the control run does not develop. Thus, the 10 m wind field remains uniform throughout. There is no banding in the reflectivity at 0600 UTC (Figure 14(c)) or 1800 UTC (Figure 15(c)).

### 6.4. No orography

The trough and vortex develop over water (Figure 13(d)), as in the control run. The isobars are more meridionally oriented at the upwind end of the English Channel, so there are subtle differences from the control run in the location and orientation of the convergence zones (Figure 13(d)) and reflectivity bands (Figure 14(d)). However, the essential pattern of air from southern England meeting air from northern France and forming the vortex is much the same.

Twelve hours later, the trough is more uniform and the wind field evolves into a single confluence zone (Figure 15(d)), as in the control run. The isobars are still more meridionally oriented, however, leading to stronger flow off the north coast of France than in the control run. The band is further north, over the south coast of England, and more cellular than in the control run, much less closely resembling the radar image (Figure 15(a)), but is of similar intensity to both.

### 6.5. Constant $z_0$

The trough develops over water, but the air stream across the east coast of England turns less markedly towards it (Figure 13(e)). Due to constant roughness length over the domain, the winds no longer accelerate over water (notice the faster winds over land than in the other panels of Figure 13). However, the distribution and intensity of convergence (Figure 13(e)) and reflectivity (Figure 14(e)) are similar to the control and no-orography runs.

Twelve hours later, the reflectivity is primarily over the south coast of England as in the no-orography run (compare Figure 15(d) and (e)). However, the band in the constant-$z_0$ run is not so concentrated over the south coast of England and weak multibands are over the English Channel.

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Table 6. A summary of all simulations, indicating whether each of orography, $z_0$ variations and surface heat and moisture fluxes was left active (Y) or turned off (N).

<table>
<thead>
<tr>
<th>Simulation</th>
<th>Orog.</th>
<th>$z_0$ var.</th>
<th>Sfc. fluxes</th>
</tr>
</thead>
<tbody>
<tr>
<td>Control</td>
<td>Y</td>
<td>Y</td>
<td>Y</td>
</tr>
<tr>
<td>No surface fluxes</td>
<td>Y</td>
<td>Y</td>
<td>N</td>
</tr>
<tr>
<td>No orography</td>
<td>N</td>
<td>Y</td>
<td>Y</td>
</tr>
<tr>
<td>Constant $z_0$</td>
<td>N</td>
<td>N</td>
<td>Y</td>
</tr>
<tr>
<td>All removed</td>
<td>N</td>
<td>N</td>
<td>N</td>
</tr>
</tbody>
</table>

Figure 12. The three domains used for all model simulations.

Figure 13. The orientation of the isobars (a) is similar to that on radar (b), as is the confluence of the two air streams. Greater wind speed over land than in the other panels of Figure 13 (c). The 10 m wind field (d) is more meridionally oriented at the upwind end of the English Channel, so there are subtle differences from the control run in the location and orientation of the convergence zones (e) and reflectivity bands (f). However, the essential pattern of air from southern England meeting air from northern France and forming the vortex is much the same. Twelve hours later, the trough is more uniform and the wind field evolves into a single confluence zone (g), as in the control run. The isobars are still more meridionally oriented, however, leading to stronger flow off the north coast of France than in the control run. The band is further north, over the south coast of England, and more cellular than in the control run, much less closely resembling the radar image (h), but is of similar intensity to both.

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6.6. All removed

As in section 6.3, the removal of surface fluxes prevents convergence (Figure 13(f)) and organized precipitation (Figures 14(f) and 15(f)) from forming. The main difference between this run and the no-surface-fluxes run (in which orography and frictional variations remained) is that there is no turning of the winds towards low pressure to the south as they cross the east coast of England. However, this is immaterial because, as demonstrated by the no-surface-fluxes run, this effect alone is insufficient to generate convergence over the English Channel.

6.7. Later initialization times

The same simulations were performed with various other initialization times (not shown). In control runs initialized later, the location and morphology of the band in its organized stage much more closely resembled the radar imagery than in Figure 15(b). However, the corresponding sensitivity experiments differed little from the control run, due to insufficient spin-up time. Thus, because the purpose of this article is to investigate the physics of the snowbands rather than to produce the most accurate simulation possible, we have presented the simulations with sufficient spin-up time for each sensitivity simulation to evolve distinctly. We maintain that the larger differences between the observed and modelled bands with the earlier initialization time do not change the conclusions that can be drawn from these simulations. Instead, we have attempted to provide an overview of the identified physical processes (orography, $z_0$ variations, surface fluxes) involved in band formation.

6.8. Irish Sea

The band documented over the Irish Sea in section 3.2 was also simulated with the same domains (but without the

Figure 13. Observations (a) and model output (b–f) at 0600 UTC on 30 November (42 h into the simulation). (a) shows Nimrod precipitation rate (mm h$^{-1}$) and wind vectors taken from Met Office Integrated Data Archive System (MIDAS) land surface stations (m s$^{-1}$). (b)–(f) show sea-level pressure (grey contours every 1 hPa, high pressure to the north), 10 m wind divergence (m s$^{-1}$ km$^{-1}$, negative values shaded), and 10 m wind vectors (m s$^{-1}$). (b) Control run; (c) no surface fluxes; (d) no orography; (e) no orography and constant $z_0$; (f) all removed. Domain as in Figure 3.
inner 1 km domain), similar spin-up time and equivalent settings (not shown). A full-physics run produced mid-sea convergence and organized precipitation, similar to the English Channel case. A simulation with surface fluxes switched off also matched that of the English Channel, with the surface convergence failing to form and the 10 m wind field remaining uniform.

7. Discussion

7.1. How does the wind speed affect band formation?

Bands form over open water in strong winds and cold advection. The following three examples illustrate that such bands require lower-tropospheric wind speeds of about 10 m s\(^{-1}\) or more. Firstly, observed bands formed in surface winds exceeding 10 m s\(^{-1}\) over the Baltic Sea (Andersson and Nilsson, 1990). Secondly, in idealized experiments over Lake Michigan, Hjelmfelt (1990) found that a wind speed of 10 m s\(^{-1}\) produces a band for all simulated \(\Delta T\) and lapse rates but smaller wind speeds only produce a band for certain \(\Delta T\) and lapse rates. Thirdly, in an idealized elongated body of water, Alestalo and Savijärvi (1985) used a wind speed of 10 m s\(^{-1}\) to produce a single band. In the present study, although statistical tests did not reveal a critical wind speed for either sea, the 850 hPa wind speed was above 10 m s\(^{-1}\) for 9 of the 10 times at which bands were observed in along-channel flow over the English Channel. For the Irish Sea, the 850 hPa wind speed was above 10 m s\(^{-1}\) at only 5 of the 10 times but was 6.7 m s\(^{-1}\) or greater at 9 of the 10 times. For both seas, the 850 hPa wind speed was significantly greater when bands were present than when they were not present. In particular, the band documented over the Irish Sea developed in an 850 hPa wind speed of 9.8 m s\(^{-1}\) but decayed when the wind speed dropped below 4 m s\(^{-1}\).

Rather than wind speed, the ratio of ambient wind speed \(U\) to fetch length \(L\) determines lake-effect morphology for an idealized circular lake (Laird et al., 2003a). When \(U/L\) is less than 0.02 m s\(^{-1}\) km\(^{-1}\), mesoscale vortices form. When \(U/L\) is between 0.02 and 0.09 m s\(^{-1}\) km\(^{-1}\), bands form. For \(U/L\) greater than 0.09 m s\(^{-1}\) km\(^{-1}\), lake-effect convection takes the form of widespread coverage. For elliptical lakes

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Figure 14. As Figure 13, but showing simulated precipitation rate (mm h\(^{-1}\), shaded) rather than 10 m wind divergence in (b)–(f).
with an aspect ratio of 9:1 (roughly that of the English Channel and Irish Sea) and flow parallel to the long axis, the threshold for bands is lowered to 0.017 (Laird et al., 2003b). Given that the English Channel and Irish Sea are both roughly 500 km long, $U/L = 0.09$ corresponds to 45 m s$^{-1}$. This value is rarely exceeded at low levels over the UK and indeed widespread coverage has not been documented over the English Channel and Irish Sea. However, $U/L = 0.017$ corresponds to 8.5 m s$^{-1}$. Thus, there is a rough consistency between the observational results in this study and the criteria in the Laird et al. studies. Additional observational and modelling studies could further refine these criteria for the waters around the UK.

7.2. How does the lower boundary condition affect band formation?

Previous studies of similar bands have discussed the relative importance of orography and land–sea frictional differences. For example, Andersson and Gustafsson (1994) found that coastal orography was of secondary importance to forming snowbands over the Baltic Sea, and Onton and Steenburgh (2001) found that the orography was not responsible for band formation over the Great Salt Lake but merely altered the distribution of the snowfall. In our simulations of the snowband over the English Channel, the removal of orography resulted in the band being composed of more cellular convection but did not alter its existence. Even without the orography, some authors have found land–sea frictional differences to be of primary importance in forming convergence for quasi-stationary bands over water. For example, using an idealized hydrostatic model, Alestalo and Savij¨arvi (1985) found that a geostrophic wind of 10 m s$^{-1}$ along the major axis of an elongated body of water was sufficient to induce convergence due to frictional difference alone, even when the land and sea are the same temperature and the atmosphere is neutrally stratified. Roeloffzen et al. (1986) also found frictional convergence alone to be sufficient to produce a quasi-stationary cloud band parallel to an idealized coastline. In our simulations, setting the roughness length $z_0$ as a constant caused winds crossing the east coast of England to turn less towards the trough over the English Channel than in the control run. The constant-$z_0$ run also produced multibands at a later...
time, with a less well-defined reflectivity maximum than in the control and no-orography runs. Thus, our simulations show that frictional differences enhance band intensity but their absence does not preclude band formation.

7.3. Are the bands caused by land breezes?

Given that the classic land breeze is a circulation resulting from the diurnal variation in heating under relatively benign synoptic situations, particularly during the warm season, we question its applicability to these snowbands. Our interpretation is based upon the following evidence.

- No diurnal cycle in the initiation of the observed snowbands was evident.
- The land–sea temperature difference changed little during the band’s development and decay (the temperature difference gradually increased; see Table 3), implying that the band was not diurnally driven.

Some authors have interpreted the land-breeze concept more broadly as a thermally driven offshore flow superimposed on an along-shore ambient wind, free from any diurnal influence, and have attributed along-shore snowbands to the collision of these land breezes from opposing shores (e.g. Passarelli and Braham, 1981; Savijärvi, 2012). In accordance with this relaxed definition, the American Meteorological Society (AMS) Glossary states that the land breeze ‘usually’ blows at night (http://amsglossary.allenpress.com/glossary).

In our simulation of the band over the English Channel, there was indeed strong offshore flow across the north coast of France that failed to develop in the absence of surface fluxes and was essentially the same in the no-orography and constant-20 runs. Thus, this aspect to the flow was largely thermally driven and qualifies as a land breeze as defined by the above authors and in the AMS Glossary. However, there was no major difference in the flow from southern England in the control run and no-surface-fluxes run.

This asymmetry between the two coasts may be explained by Atlas et al. (1983). They found that, during cold-air outbreaks, a bay that is concave in the downwind direction leads to offshore convergence downwind because the part of the air stream that has had a shorter trajectory over the warm water experiences less heating than the other trajectories. Thus, there is differential heating and the higher pressure air that has travelled less distance over water flows towards the lower pressure air further offshore. For this case, the north coast of France is divided into two such bays that are concave facing downwind (east and west of the Cherbourg Peninsula, Figure 2) and along each of these there is strong turning of the winds towards the air that has travelled further over water (to the right). This was also observed over the English Channel in the modelling experiments of Monk (1987). The south coast of England, on the other hand, is almost a straight line, which does not allow for this effect.

Combining the evidence from sections 7.1, 7.2 and 7.3, the snowband is the mesoscale response to surface heating of cold-air advection over warm water, as first described by Lavoie (1972). The roles of orography and frictional differences may affect the timing or location of the band but do not fundamentally alter its occurrence.

8. Conclusions

Quasi-stationary wind-parallel snowbands formed along the major axis of the English Channel and Irish Sea during cold-air outbreaks in the winters of 2009–10 and 2010–11. These bands bear similarity to those in other parts of the world: Great Lakes, Great Salt Lake, New England, off the north coast of Germany, Sea of Japan and the Baltic Sea. They were studied using the analysis of observational data (e.g. radar imagery, satellite imagery, surface and upper-air data), synoptic composites, climatology and real-data numerical modelling. Two different morphologies of bands were identified: single bands and multibands. Unfortunately, the climatology in section 5 was unable to identify distinguishing characteristics between these two morphologies, suggesting that the environmental factors tested in this study (e.g. wind direction, land–sea temperature difference) were not responsible for the single or multiband morphology. However, case studies over both seas in section 3 indicated that multibands occurred during the formation and decay stages of mature single-band cases, which was commonly (although not always) observed during the winters in question.

The case studies showed that bands formed over the English Channel and Irish Sea when winds were greater than about 10 m s$^{-1}$ and nearly parallel to the long axis of the water body. The surface-to-850 hPa temperature difference exceeded 18 K and the land–sea temperature difference exceeded 10 K. The band over the English Channel decayed when the wind direction was no longer along-channel. The band over the Irish Sea decayed when the wind speed decreased markedly.

Synoptic composites showed that bands formed with a ridge over Greenland and a trough over western Europe at 500 hPa. A surface anticyclone was centred west of Iceland and a zonally elongated cyclone was centred over western Europe. Synoptic composites were similar, however, for times at which bands were not present, demonstrating that band occurrence depends on the finer details of the flow environment. In support of this, surface-to-850 hPa temperature difference and 850 hPa wind speed were both found to be significantly greater at times of banding than for no banding. Thus the observed bands formed when convection resulted from pronounced instability over water, with that convection organized into bands for sufficiently strong winds.

Model sensitivity experiments were performed for one such band over the English Channel. In the control run, a trough in sea-level pressure formed over the English Channel and surface winds turned cyclonically towards the trough as they passed from the North Sea over the east of England. Over the English Channel, these winds met an air stream offshore from northern France, forming a mesoscale vortex and strong convergence zones over the English Channel. Precipitation bands formed over the convergence lines. Over the course of about 12 h, the trough intensified, forming a single convergence line and precipitation band at the downwind end.

Removing the orography and setting the roughness length constant across land and sea produced a weaker trough and differences in the timing, location and intensity of the band, but the band was still present, despite less turning of the winds over southern England towards the trough. The removal of surface heat and moisture fluxes, both with
and without orography and $z_0$ variations, resulted in the complete failure of convergence and organized precipitation to develop. We therefore conclude that these fluxes are the major factor in generating the observed snowbands.

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The Nimrod data used in this study were provided by the Met Office through the British Atmospheric Data Centre (BADC). ECMWF analyses were obtained using the ECMWF Meteorological Archive and Retrieval System (MARS) software. Sounding data were provided mainly by the Department of Atmospheric Science, University of Wyoming, but some were only available through BADC. Meteosat images and MIDAS land-surface-station data were also accessed through BADC. We thank Karen Barfoot of the Met Office for providing SST data, NOAA/ESRL Physical Sciences Division for construction of the composites in Figure 11 from their web page and Bogdan Antonescu and Hugo Ricketts for use of their script to process the raw Nimrod data. We also thank the anonymous reviewers and Keith Browning, who all provided helpful comments on the initial version of the manuscript. Jesse Norris is a NERC-funded student through the DIAMET (DIAbatic influences on Mesoscale structures in ExTratropical storms) project, NE/I005234/1.

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6 Summary and conclusions

This thesis has investigated the sensitivity of various precipitation bands in the midlatitudes to diabatic influences: surface friction, sea-surface temperature, and, in Papers 1 and 2, latent-heat release. The sensitivity of bands has been assessed via a climatology of precipitation-radar, sounding, and SST data in Paper 3, but throughout the thesis by simulations with the WRF model. Paper 1 took a uniform initial jet and zonally homogeneous lower boundary to investigate which bands form within a baroclinic wave without any of the complications in the initial and boundary conditions that may affect bands in the real atmosphere, and how sensitive they are to the diabatic processes. Paper 2 employed the same methodology of Paper 1, but at higher resolution and focusing on precipitation cores along the cold front. Paper 3 documented snowbands over the English Channel and Irish Sea in the winters of 2009–10 and 2010–11, which may be considered extreme instances of postfrontal bands within baroclinic waves, with an observational climatology of the bands and WRF sensitivity simulations of one band over the English Channel.

To summarise, the three papers have made the following findings:

6.1 Paper 1

Idealised baroclinic-wave simulations were performed at 20-km grid spacing to investigate the formation and sensitivity of different types of precipitation bands, without the complications of nonuniform initial conditions or complexities in the lower boundary. Surface friction was found to slow the cyclone’s development and enhance the cold front, relative to the warm front, as previous studies have found. The new element is that friction enhances the narrow rainband in baroclinic-wave simulations and horizontally separates it from the wide warm-frontal band. Latent-heat release and surface heat fluxes enhanced and hindered multiple banding, respectively, in the mature stage of the cyclone, when the cold front transitioned from ana- to kata-type. Latent-heat release narrowed and intensified separate mid-tropospheric vertical-velocity maxima within the synoptic-scale ascent along the warm conveyor belt, resulting in multiple warm-sector bands. Surface fluxes allowed the ana cold front to persist longer, by preventing the protrusion of the tropopause fold above the surface cold front, so that the warm-sector precipitation remained tied to the cold front and the multiple bands did not form.
6.2 Paper 2

A simulation from Paper 1 was re-initialised, after the formation of the cold front, and a nested domain of 4-km grid spacing was inserted to apply the same methodology of Paper 1 to the finer-scale precipitation cores along the cold front. In the control simulation, the precipitation cores were shorter and at a greater angle to the cold front, hence less closely resembling a continuous rainband, at the equatorward than poleward end of the cold front. Latent-heat release and surface friction were both found to enhance the multiple cores, while greater surface fluxes inhibited multiple cores. In the absence of latent heat, the cores were shorter and closer together, with no distinctive gaps forming between cores. In simulations with roughness lengths appropriate to highly rough land surfaces, the most distinctive and longest precipitation cores formed, such as have been observed by precipitation radar over land, at large angles to the cold front with large gaps between cores. Greater SST, on the other hand, despite enhancing the width and intensity of the cold-frontal rainband, caused the cores to form a continuous rainband more quickly, with no cross-frontal precipitation maxima.

6.3 Paper 3

Extreme postfrontal snowbands in the winters of 2009–10 and 2010–11 over the English Channel and Irish Sea were investigated, via a climatology of sounding and SST data, and real-data simulations of one such band over the English Channel. The climatology indicated that these bands were favoured by greater lower-tropospheric lapse rate (hence greater surface fluxes), stronger lower-tropospheric winds parallel to the relevant body of water’s major axis, and, for the English Channel, greater land–sea temperature contrast (hence differential surface fluxes between land and sea). The real-data simulations confirmed the importance of the surface fluxes on the snowbands, in which the band failed to form when the surface fluxes were switched off. The topography around the English Channel (land–sea frictional variations and orography) helped organise the wind field into the single smooth convergence line that made the band so well defined. Therefore, these factors significantly contributed to the band’s formation and morphology, but, unlike surface fluxes, were not essential for the formation of banded precipitation.
6.4 Land versus sea bands

A common element to all three papers is the contrast between bands over land and over sea. In Paper 3, this contrast was investigated explicitly, via the real-data simulations. In the idealised simulations in Papers 1 and 2, this contrast was investigated implicitly, via experiments of sensitivity to surface friction, latent-heat release, and surface fluxes.

Over land, surface friction is greater and, particularly in wintertime, as in Paper 3, we may expect that surface fluxes are less than over sea. Because of the reduced moisture flux from land surfaces, we may also expect less latent heating and cooling over land. These differences between land and sea may affect whether bands are single- or multiply-banded (section 6.4.1), and also the intensity of bands forced at the surface (section 6.4.2).

6.4.1 Single versus multiple bands

Papers 1 and 2 both found that high friction, as over land, may enhance multiple banding, in the across-frontal (Paper 1) and along-frontal (Paper 2) directions, at the two different horizontal scales. Low surface fluxes, as there may be over land compared to sea, also enhanced multiple banding in both the along-frontal and across-frontal directions. Therefore, multiple banding may be doubly favoured over land by these two factors. Figure 1 of Paper 2 supports this hypothesis, where the precipitation cores are more distinctive over land than over sea in all four of the radar composites.

However, in both Papers 1 and 2, latent-heat release also enhanced multiple banding. Therefore, if latent-heat release is reduced over land, the enhancement of multiple bands by high friction and low surface fluxes may be less than it would otherwise be.

Paper 3 also investigated why sometimes single and sometimes multiple bands formed. The climatology did not reveal any significant differences in the flow or stratification between times of single and multiple banding, but multiple banding was commonly observed on radar to occur during the formative and decay phases of the bands. The WRF simulation of a snowband also exhibited this evolution: Multiple bands were present while the wind field was poorly organised, with multiple offshore convergence lines arising from the complexities of the coastlines. The wind field then smoothed over about twelve hours into a single almost perfectly straight convergence line, less defined by the complexities of the coastlines. Thus, this evo-
olution suggests that complex coastlines favour multiple bands, while a more uniform channel would be more likely to produce single bands. This distinction may also be true for other bands forced at the surface, e.g., precipitation cores along the cold front (see section 6.5).

6.4.2 Intensity of surface-forced bands

The results of this thesis also suggest that the intensity of bands forced at the surface differ between land and sea. In Papers 1 and 2, in simulations with greater friction, the precipitation along the surface cold front, although taking longer to intensify, was eventually more intense. Greater surface fluxes also intensified the precipitation along the surface cold front in both Papers 1 and 2. Therefore, over land, if surface fluxes are less than over sea, then friction and surface fluxes may have opposing effects in terms of the intensity of bands forced at the surface.

In Paper 3 also, the intensity of the simulated snowband was enhanced by strong surface fluxes over sea that were not present over land. Unlike Papers 1 and 2, the reduced friction over sea did not appear to inhibit the intensity of the bands. This contrast between the cold-frontal and postfrontal bands may be because the postfrontal bands in question were wind-parallel, whereas the cold-frontal bands were roughly wind-perpendicular. Therefore, for cold-frontal bands, frictional deceleration enhances convergence along the band, whereas, for the postfrontal bands, frictional deceleration enhances convergence across the band. The climatology in Paper 3 in fact suggested that the bands were favoured by strong low-level winds (hence favoured by low friction), although the previous literature suggests that strong winds (specifically, strong vertical shear) aid the organisation of postfrontal bands, rather than the intensity. Also in contrast to the bands in Papers 1 and 2, the snowbands' intensity was dependent on the contrast between land and sea, both thermal and frictional, to form such a strong convergence line.

Sections 6.4.1 and 6.4.2 illustrate that bands over land are likely to differ to bands over sea in both number and intensity. These differences may be the result of multiple contrasts between land and sea (e.g., friction, latent-heat release, and surface fluxes). In the real atmosphere, the individual effects of each of these contrasts cannot be isolated, which is what this thesis has done.
6.5 Further work

The initial further work to follow this thesis is to continue working on Paper 2, which is the only paper of the three not yet submitted to or published in a journal. As documented in Paper 2, the differences in the morphology of precipitation cores between simulations of varying roughness length, latent heat, and SST may be related to differences in the flow and stratification along and across the cold front. This issue will be further investigated to determine whether precipitation cores along the cold front in a realistic model are indeed due to horizontal-shear instability, as interpreted by previous studies of precipitation cores.

More generally, the work to follow this thesis consists of further sensitivity studies of precipitation bands in idealised-baroclinic-wave simulations. In the idealised-modelling experiments in this thesis, three major simplifications were made to the baroclinic-wave simulations, which are not representative of the real atmosphere:

1. As argued in the introduction, the initial conditions for an idealised baroclinic wave may profoundly alter the structure of precipitation bands within the baroclinic wave. There was no horizontal shear in the initial jet used for this thesis, which can profoundly alter the structure of a cyclone forming on the jet and of the associated fronts (section 1.1.4). Therefore, bands forming in baroclinic-wave models with initial cyclonic or anticyclonic shear (or, following Schultz et al. 1998, confluence or diffluence), may be fundamentally different to those simulated in this thesis and form in different cyclone-relative locations. As discussed in the introduction, Gray and Dacre (2008) investigated the sensitivity of rainbands to the initial conditions in an idealised deformation-strain model. Further studies should expand the work of Gray and Dacre and perform multiple baroclinic-wave simulations with different initial conditions, as in, e.g., Davies et al. (1991), Thornicroft et al. (1993), and Schultz et al. (1998), but at higher resolution (20-km grid spacing or less) to investigate how the distribution and morphology of precipitation bands vary. The intensity of the initial jet may also be changed to investigate the sensitivity of cyclones forming on the jet and of the bands within the cyclones.

2. Orography has been deliberately excluded from the idealised simulations in this thesis, and only investigated in the real-data simulations to check that orography was not a crucial factor in forming the observed bands. However, in the real atmosphere, orography may enhance bands within extratropical cyclones (e.g., Braun et al. 1997; Doyle 1997; Miniscloux et al. 2001; Kirshbaum
and Durran 2005; Viale et al. 2013; Paper 3 of this thesis) and even generate bands that otherwise would not exist (e.g., Kirshbaum et al. 2007; Schumacher et al. 2010; Barrett et al. 2013). Further idealised baroclinic-wave simulations should be performed to investigate the effect of orography on the bands simulated in Papers 1 and 2. For example, a land mass could be created, such as in Paper 2, but containing an idealised mountain range. The effect on the bands (either the larger-scale bands in Paper 1 or the finer-scale precipitation cores in Paper 2) would be readily evident by a comparison between simulations with and without orography. Furthermore, any bands forming in a simulation with orography that do not form in an otherwise-equivalent simulation without orography, i.e., in different cyclone-relative locations to those simulated in Paper 1, would showcase the ability of orography to produce bands in extratropical cyclones.

3. Diurnal effects were also removed, by the exclusion of radiation from the simulations. Over the ocean, cyclones and fronts are relatively unaffected by the diurnal cycle (e.g., Reeder 1986; Physick 1988; Garratt et al. 1989). However, over land, daytime heating of the boundary layer can be frontolytical, due to turbulent stresses retarding the surface winds, and nocturnal cooling of the boundary layer can be frontogenetical, due to the weakening of that turbulence (e.g., Pagowski and Taylor 1998; Reeder and Tory 2005; Thomsen et al. 2009).

Where a coastline exists with diurnal effects, cold fronts aligned perpendicular to the coastline and moving along the coastline in the daytime may develop a bulge, with the part of the front over land moving more slowly than that over water (e.g., Garratt 1986; Gallus and Segal 1999). This effect may be partly explained by gravity-current theory, due to the greater sensible heat flux over land. By contrast, cold fronts aligned parallel to a coastline and approaching the coastline in the daytime may be enhanced by the co-existing sea-breeze front (e.g., Physick 1988; Brümmer 1995; Rhodin 1995). Muir and Reeder (2010) investigated this effect via idealised modelling and found that cold fronts approaching a coastline in the afternoon, when sensible heating over land is maximised, become even stronger than the superposition of the cold front and sea-breeze front, and surge inland in the evening, when the turbulent mixing subsides.

We may expect that these diurnal effects on the intensification and motion
of fronts also have profound influences on the bands that form along and near fronts. This hypothesis could be investigated by further baroclinic-wave simulations with the inclusion of a coastline, and radiation and land-surface parameterisation.

The investigation of the snowbands in Paper 3 also raises further questions. This study was limited to the winters of 2009–10 and 2010–11 in which the motivating snowbands occurred. Although the climatology identified favourable lower-tropospheric flow and stratification characteristics, and the simulations identified the sensitivity of one particular band to the various diabatic influences, the conclusions of the study were limited to these bands over the English Channel and Irish Sea. To generalise the sensitivity of postfrontal bands to land–sea contrasts, idealised simulations should be performed, whether with a baroclinic-wave or a simpler model, with an elongated body of water, surrounded by land, below a postfrontal air mass. Sensitivity experiments, varying the geometry of the body of water, and the thermal and frictional contrasts between land and sea could determine the generality of the snowbands studied in Paper 3.

Idealised-modelling studies along these lines have been done by Laird et al. (2003a,b), as described in Paper 3. However, these studies only investigated bands forming over circular and oval-shaped bodies of water. As argued in section 6.4.1, the complexities in real coastlines may significantly alter the morphology of bands, e.g., single versus multiple bands, forming over a body of water. Therefore, the insertion of coastline complexities, such as those that exist around the English Channel and Irish Sea, into such simulations may produce fundamentally different morphologies of bands. Such simulations could also be performed for other flow regimes to establish whether multiple banding is favoured by complex coastlines for other band types, e.g., precipitation cores along the cold front.
Figure 21: Original version of Fig. 11 of Paper 3 of this thesis.
Figure 22: Original version of Fig. 13 of Paper 3 of this thesis.
Figure 23: Original version of Fig. 14 of Paper 3 of this thesis.
Figure 24: Original version of Fig. 15 of Paper 3 of this thesis.
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