Variability of Precipitation along Cold Fronts in Idealized Baroclinic Waves

JESSE NORRIS, GERAINT VAUGHAN, AND DAVID M. SCHULTZ
Centre for Atmospheric Science, School of Earth, Atmospheric and Environmental Sciences, University of Manchester, Manchester, United Kingdom

(Manuscript received 25 October 2016, in final form 13 February 2017)

ABSTRACT
Precipitation patterns along cold fronts can exhibit a variety of morphologies including narrow cold-frontal rainbands and core-and-gap structures. A three-dimensional primitive equation model is used to investigate alongfront variability of precipitation in an idealized baroclinic wave. Along the poleward part of the cold front, a narrow line of precipitation develops. Along the equatorward part of the cold front, precipitation cores and gaps form. The difference between the two evolutions is due to differences in the orientation of vertical shear near the front in the lower troposphere: at the poleward end the along-frontal shear is dominant and the front is in near-thermal wind balance, while at the equatorward end the cross-frontal shear is almost as large. At the poleward end, the thermal structure remains erect with the front well defined up to the mid-troposphere, hence updrafts remain erect and precipitation falls in a continuous line along the front. At the equatorward end, the cores form as undulations appear in both the prefrontal and postfrontal lighter precipitation, associated with vorticity maxima moving along the front on either side. Cross-frontal winds aloft tilt updrafts, so that some precipitation falls ahead of the surface cold front, forming the cores. Sensitivity simulations are also presented in which SST and roughness length are varied between simulations. Larger SST reduces cross-frontal winds aloft and leads to a more continuous rainband. Larger roughness length destroys the surface wind shift and thermal gradient, allowing mesovortices to dominate the precipitation distribution, leading to distinctive and irregularly shaped, quasi-regularly spaced precipitation maxima.

1. Introduction
Precipitation patterns along cold fronts as observed by radar can exhibit a variety of morphologies. Sometimes the precipitation can fall within a single narrow line called a narrow cold-frontal rainband (e.g., Browning and Harrolld 1970; Hobbs 1978; Houze and Hobbs 1982; Knight and Hobbs 1988). Other times, the precipitation can break up into regularly spaced cores of maximum precipitation rate separated by gaps of lighter or no precipitation in between, called core-and-gap structures (Hobbs and Biswas 1979; James and Browning 1979; Hobbs and Persson 1982). Often, the cores are rotated anticyclonically from the orientation of the surface cold front. These features are clearly observed over land (e.g., over the British Isles as in Fig. 1) and have also been observed over open ocean (e.g., Hobbs and Biswas 1979; Hobbs and Persson 1982; Wakimoto and Bosart 2000; Jorgensen et al. 2003). The precipitation associated with these fronts is often shallow, with hydrometeors only reaching heights of 2–4 km (James and Browning 1979; Hobbs and Biswas 1979; Hagen 1992; Browning and Reynolds 1994; Brown et al. 1999; Wakimoto and Bosart 2000; Jorgensen et al. 2003). As well as introducing pronounced local variability in rainfall accumulation, the circulations associated with

 DOI: 10.1175/MWR-D-16-0409.1
© 2017 American Meteorological Society
FIG. 1. Met Office precipitation-radar composites at 1-km grid spacing, expressed as precipitation rate in mm h\(^{-1}\), during the passage of cold fronts over the British Isles at (a) 1600 UTC 29 Nov 2011, (b) 1350 UTC 2 Nov 2013, (c) 1215 UTC 11 Nov 2010, (d) 2000 UTC 18 Dec 2013, and (e) 1800 UTC 22 Nov 2012. Boxes are drawn in (a)–(d) to highlight finescale precipitation cores, but the larger-scale precipitation cores in (a) over southern England should be self-evident.
precipitation cores can spawn tornadoes (e.g., Mulder and Schultz 2015; Apsley et al. 2016).

A considerable variety of morphologies of precipitation cores exists from case to case. Cores may be long and spaced far apart (Fig. 1a) or short and spaced much more closely together (Fig. 1b). Cores may also be more curved in some cases than in others (Figs. 1c,d; in both these images the cores point into the prefrontal air mass, likely associated with mesovortices along the front). Furthermore, cores may vary within a single case, with their length, width, spacing, curvature, and cold-front-relative angle varying along the cold front (particularly exhibited in the boxes drawn in Figs. 1a and 1c). Cores may even be embedded within a wide band along the cold front (Fig. 1d). In contrast, some cold fronts maintain an almost continuous line of maximum precipitation (Fig. 1e). Given this wide range of possible morphologies, the natural questions are what are the factors that cause these different morphologies of precipitation core-and-gap regions along cold fronts, and why do some fronts not develop the core-and-gap morphology?

Cores and gaps have commonly been proposed to result from horizontal shear instability, for which many studies have gained observational evidence (e.g., Matejka 1980; Carbone 1982; Hobbs and Persson 1982; Browning and Roberts 1996). Also, some studies have performed real-data simulations of cores and gaps, 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). This paper seeks to build on these previous studies by addressing the question of why some fronts develop cores and gaps and some do not, in terms of the synoptic environment.

When the cloud-layer wind is oriented along a front, precipitation tends to fall parallel to the front, whereas fronts with a large component of cloud-layer wind normal to the front are more likely to generate precipitation maxima pointing away from the front (Dial et al. 2010). Jorgensen et al. (2003) made detailed in situ measurements of the flow environment along and across a cold front with cores and gaps over the eastern Pacific. They found that the vertical shear of the low-level cross-frontal wind was closely linked to the resulting precipitation structures. At the poleward ends of cores, where the precipitation maxima were farthest displaced from the general orientation of the cold front, the cross-frontal vertical shear was greatest and the updraft erect or downshear tilted (i.e., eastward). At the equatorward ends of cores and in gaps, where the precipitation maxima were closest to the cold front, the cross-frontal shear was weaker with the updraft broader and upshear tilted (up the cold front, see Fig. 25 in Jorgensen et al. for a schematic).

The above studies suggest that the development of vertical shear across a cold front applies a perturbation to the wind field above the surface cold front and results in cyclonically oriented precipitation cores. In contrast, a cold front with more alongfront-oriented vertical shear should lead to a more uniform wind field along the front and be host to more uniform rainfall. These two situations may be expected to occur in the event of kata-type and ana-type cold fronts, respectively. An ana-type cold front is characterized by rearward sloping ascent from the warm conveyor belt up the cold front, so that the precipitation is aligned along the cold front (e.g., Browning and Pardoe 1973). A kata-type cold front is crossed by an intrusion of dry air from upper levels, so that an upper-level front forms above and ahead of the surface cold front and some precipitation falls ahead of the cold front (e.g., Browning and Monk 1982).

To isolate the physics associated with the variation in the morphology of precipitation cores, idealized simulations have been performed in previous studies. Bluestein and Weisman (2000) showed that vertical shear 45° from the frontal orientation is most likely to generate regularly spaced and shaped cells along the front (such as in Fig. 1b), whereas vertical shear normal to the front is more likely to generate larger isolated cells that are irregularly spaced and shaped (such as in Fig. 1c). However, these structures are sensitive to more than just the shear environment. Kawashima (2007) simulated precipitation structures resembling observed cores and gaps from a wavelike disturbance that gained its energy from both vertical shear and buoyancy along the cold front, demonstrating that more than one mechanism may produce such structures. However, Kawashima (2011) noted that the simulations of Kawashima (2007) were initialized with weak vertical shear and only static stability was varied between simulations, hence were restricted in their generality and applicability to fronts in the real atmosphere. To achieve greater generality, Kawashima (2011) simulated precipitation structures along a cold front resembling observed precipitation cores, varying the vertical shear along/across the front, the magnitude of the wind shift across the front, and the prefrontal static stability between simulations. The cores and gaps he simulated were sensitive to all of these factors, suggesting that cores and gaps in the real atmosphere result from the nonlinear interaction of these factors.

These idealized-modeling studies demonstrate how precipitation cores are sensitive to the flow and stratification in the vicinity of the cold front along which they form. However, in the real atmosphere, the flow and stratification along a cold front may vary considerably,
leading to a range of morphologies of cores along the front. To this effect, Browning and Roberts (1996) documented a cold front over the British Isles whose precipitation morphology was markedly different between the two ends of the cold front. At the equatorward end, an ana-type cold front led to relatively uniform precipitation along the cold front, with regularly spaced cores and small spaces between cores. At the poleward end, a kata-type cold front led to more poorly organized cores with greater spaces between.

In this paper, we investigate how the synoptic environment determines which of the many possible morphologies the cores and gaps can adopt. The study builds on that of Norris et al. (2014) who compared the distribution and evolution of precipitation bands in idealized baroclinic simulations of the atmosphere with roughness length, latent-heat release, and surface fluxes of sensible and latent heat varied. Norris et al. (2014) simulated, at 20-km grid spacing, precipitation bands resembling those observed in the real atmosphere that were sensitive to all of these influences. The variation of the bands between simulations occurred via variations in the synoptic- and mesoscale structure of the flow environment. The current study aims to do the same, but for finer-scale precipitation cores. Therefore, a simulation from Norris et al. (2014) with all these diabatic factors appropriate to an extratropical cyclone over the open ocean, after 132 h when the surface cold front has formed, is reinitialized with a 4-km nested domain and the different diabatic factors are varied between simulations, subsequent to this time. Thus, this paper demonstrates how precipitation cores along a mature cold front of maritime origin vary, depending on differences in the synoptic and mesoscale flow environment.

2. Method

Moist idealized baroclinic-wave simulations were performed with version 3.7.1 of the Advanced Research Weather Research and Forecasting (WRF) Model (Skamarock et al. 2008). As in Norris et al. (2014), WRF was initialized with the baroclinic-wave test case, which consists of a zonal jet on an $f$ plane ($f = 10^{-4} \text{s}^{-1}$) in thermal wind balance with a horizontal temperature gradient at the surface of roughly $20 \text{K} (1000 \text{km})^{-1}$. The jet is obtained by inverting a baroclinically unstable potential vorticity distribution in the $y$–$z$ plane, as in Rotunno et al. (1994). The computational domain is 8000 km in the north–south direction. In the east–west direction, the domain is 4000 km long, which is equal to the wavelength of the most unstable normal mode of the initial jet (Plougonven and Snyder 2007). The domain has 20-km grid spacing, with 80 vertical levels from the surface up to 16 km. The lower boundary condition is ocean with a roughness length $z_0 = 0.2 \text{mm}$ and the sea surface temperature (SST) is fixed to the initial temperature of the lowest atmospheric model level. The simulations use the following parameterizations: Thompson et al. (2008) microphysics, Kain–Fritsch convection (Kain and Fritsch 1990; Kain 2004), MM5 surface layer (Monin and Obukhov 1954; Skamarock et al. 2008), and Yonsei University boundary layer (Hong et al. 2006).

After 132 h into this simulation, a wide band of precipitation lies along and ahead of a well-defined surface cold front and wind shift (Fig. 2), similar to composite real cyclones (e.g., Fig. 8 in Field and Wood 2007) and idealized modeled cyclones (e.g., Fig. 4 in Zhang et al. 2007). In the present paper, we focus on the precipitation at the leading edge of the cold front. A nested domain with 4-km horizontal grid spacing is inserted at the location indicated by the black box in Fig. 2, capturing the evolution of the cold front in the 36 h that it takes for the cold front to cross the nested domain. After the front begins to pass into the nested domain, several hours are needed for spinup in terms of simulating the front at higher resolution, during which an adjustment is
evident in the resolution of the front. Therefore, the output from the inner domain is only documented from 21 h after the inner domain is initialized onward, which is also the time it takes for the front and its associated precipitation to be fully inside the inner domain. In the remainder of this article, the hour of the simulation refers to the number of hours since the 4-km domain was initialized.

As the purpose of this article was to simulate the precipitation morphologies along the cold front, 4-km grid spacing is sufficient. Several of the figures in Jorgensen et al. (2003) of core-and-gap structures along a cold front were plotted from a simulation with a 4-km horizontal grid spacing, demonstrating that a grid spacing of 4 km is sufficient to capture the core-and-gap structures. Moreover, their simulations were nested down to 1.3-km grid spacing in order to calculate air-parcel trajectories on this scale. Although the structures on the 1.3-km grid possessed more detail, the overall structures were not substantially different between the two grids. As we are not interested in this level of computational detail, 4-km grid spacing is sufficient for this article.

In the control simulation (CNTL), the simulation is kept the same as the outer domain over the 132 h before the nest is initialized (i.e., this is an all-ocean simulation), except that convective parameterization is switched off in the 4-km nest. In the other simulations, the lower boundary is converted to half-ocean–half-land, with ocean and land occupying the left-hand and right-hand sides of the outer domain, respectively (the bold red line in Fig. 2 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 left-hand 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, $T_0$ (the initial temperature when the baroclinic wave is initialized, as opposed to at the initialization time in this paper). The variability in the sensitivity simulations is all in the thermal and frictional characteristics of the right-hand land side of the domain, as summarized in Table 1. The LANDFRIC1, LANDFRIC2, and LANDFRIC3 simulations contain a frictional contrast only between land and sea, with land roughness lengths of 5, 250, and 2000 mm prescribed, respectively (appropriate for featureless land, high crops, and a city center, respectively; World Meteorological Organization 2008). These roughness lengths span the full range of possible roughness of a land surface. In reality, the roughness length is constantly changing as a front passes over a land surface, but these simulations allow the effect on a cold front of three distinct land-use categories to be isolated. The MINUS2K, MINUS1K, PLUS1K, and PLUS2K simulations contain a thermal contrast only, with land surface temperatures of $T_0 - 2$, $T_0 - 1$, $T_0 + 1$, and $T_0 + 2$ K, respectively. These are fairly arbitrary choices and were determined partly because particularly large or small land surface temperatures led to increasingly unphysical-looking simulations. As will be shown, this 4-K difference in land surface temperature between simulations leads to significant differences. 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, MINUS2K, MINUS1K, PLUS1K, and PLUS2K, the lower boundary is treated as an ocean surface with a roughness-length or SST discontinuity, in order to isolate the sensitivity of precipitation cores to each of these factors. “Coastline” is used to refer to the discontinuity, whether frictional or thermal. The variations between simulations detailed in Table 1 are effective only after the simulations are reinitialized at 132 h.

### Table 1. A summary of the sensitivity simulations. Gives roughness length $z_0$ and the SST distribution ($T_0$ is the initial temperature of the lowest model level when the baroclinic wave is initialized). Boldface entries are where the given factor is different to CNTL. The given factors are prescribed on the right-hand side of the domain only (to the right of the red line in Fig. 2) and, in all simulations, the left-hand side of the domain is as in CNTL.

<table>
<thead>
<tr>
<th>Simulation</th>
<th>$z_0$ (mm)</th>
<th>SST (K)</th>
</tr>
</thead>
<tbody>
<tr>
<td>CNTL</td>
<td>0.2</td>
<td>$T_0$</td>
</tr>
<tr>
<td>LANDFRIC1</td>
<td>5</td>
<td>$T_0$</td>
</tr>
<tr>
<td>LANDFRIC2</td>
<td>250</td>
<td>$T_0$</td>
</tr>
<tr>
<td>LANDFRIC3</td>
<td>2000</td>
<td>$T_0$</td>
</tr>
<tr>
<td>MINUS2K</td>
<td>0.2</td>
<td>$T_0 - 2$</td>
</tr>
<tr>
<td>MINUS1K</td>
<td>0.2</td>
<td>$T_0 - 1$</td>
</tr>
<tr>
<td>PLUS1K</td>
<td>0.2</td>
<td>$T_0 + 1$</td>
</tr>
<tr>
<td>PLUS2K</td>
<td>0.2</td>
<td>$T_0 + 2$</td>
</tr>
</tbody>
</table>

#### 3. Evolution of precipitation cores and surface wind field in control simulation

After 21 h, the cold front is fully inside the inner domain and a near-continuous line of maximum precipitation lies along it (Fig. 3). At the poleward end, the precipitation undergoes relatively little evolution, maintaining a continuous convective line (Fig. 4) and resembling the narrow cold-frontal rainband shown on
radar in Fig. 1e. At the equatorward end, however, the initial narrow cold-frontal rainband starts to break up into core-and-gap regions from the south to the north (Fig. 5). By 34 h, the precipitation cores are well-defined and have formed appendages pointing into the warm air at their poleward ends, resembling the precipitation cores observed on radar in Figs. 1a–d.

A strip of maximum vorticity forms along the front, which remains intact throughout the simulation (Figs. 4, 5).

Unlike horizontal shear instability along the cold front, where the maximum vorticity rolls up, vorticity structures appear in the simulation ahead of the front. Initially, these are chaotic, but they become more organized and move poleward along the front as the simulation progresses, eventually creating bands of vorticity rotated clockwise with respect to the front, similar to the precipitation cores. These structures generate undulations in the lighter prefrontal precipitation at the both the poleward (Fig. 4) and equatorward (Fig. 5) ends. Although the strip of maximum vorticity along the front does not roll up, the mesovortices ahead of the front originate on the line of maximum vorticity.

Also from 22 h onward, separate vorticity maxima begin to form behind the cold front, becoming well-defined and periodically spaced at the equatorward end by 26 h (Fig. 5). These vorticity structures behind the front originate as a strip of vorticity in the low center, as shown by an animation of the outer domain, and develop equatorward as the cyclone occludes (not shown). Likely to be associated with horizontal shear instability, these postfrontal vorticity maxima propagate equatorward, the opposite direction to the prefrontal maxima, and point in the opposite direction to the prefrontal maxima. These postfrontal maxima never become connected to the poleward part of the front (Fig. 4), but they become connected to the strip of maximum vorticity along the equatorward part of the front after 28 h and generate undulations in the postfrontal precipitation (Fig. 5).

Thus, when the baroclinic wave is simulated at 4-km grid spacing, mesovortices either side of the front generate undulations in the lighter warm-frontal and postfrontal precipitation. At the equatorward end, there are both prefrontal and postfrontal vorticity structures and hence undulations in precipitation on either side of the cold front, from the interaction of which the precipitation cores appear to form. At the poleward end, the postfrontal vorticity structures are much farther behind the cold front, so that there are only undulations in the warm-frontal precipitation.

4. Diagnostics of precipitation cores

The continuity of the strip of maximum vorticity in the 10-m winds along the cold front (Figs. 4, 5) gives a useful reference point throughout the simulation. Alongfront variability may therefore be investigated relative to this well-defined frontal location by deriving diagnostics to describe the evolution of the precipitation cores and of the flow environment in which they form. Therefore, every hour, at every y coordinate in the inner domain, the corresponding x coordinate was found at which the 10-m vorticity gradient in the x direction (roughly
FIG. 4. Evolution of surface precipitation rate (colors, mm h\(^{-1}\)) and 10-m relative vorticity (contours at 5, 10, 20, 50, and \(100 \times 10^{-5}\) s\(^{-1}\)) in the poleward box.
Fig. 5. As in Fig. 4, but in the equatorward box.
cross-frontal) was greatest. For every hour of the simulation, this method generated a set of grid points from south to north, giving the western edge of the cold front. The location of the front is shown at 21 h in Fig. 3 by the leftmost bold line, demonstrating the effectiveness of this method in identifying the front. The line of maximum precipitation maximum is just ahead of the identified front, with near-constant distance between the identified front and the maximum precipitation. The poleward and equatorward boxes in Fig. 3 each stretch 90 grid points (360 km) from north to south along the front (the parts of the front shown in Figs. 4 and 5, respectively) and the state of the atmosphere above the front may be compared between these two sections.

The vertical wind profile above this identified front is markedly different between the poleward and equatorward ends (Fig. 6). The hodographs are calculated from the mean wind along the front at the equatorward and poleward ends. The hodographs shown are at 21 h, before the cores and gaps form in the rainband at the equatorward end, but the essential patterns remain the same throughout the period shown in Figs. 4 and 5. At both the equatorward and poleward ends, winds along the surface front are northwesterly, becoming southwesterly just above the surface, with positive vertical shear of zonal and meridional wind up to 400 hPa. This shear is greater at the poleward end, particularly in the $v$ component, where shear is almost purely meridional (roughly along-frontal) from about 950 to 850 hPa, with a magnitude of about 30 m s$^{-1}$ difference in this layer. Contrastingly, at the equatorward end, this layer of along-frontal shear is just between about 950 and 900 hPa with a magnitude of 15 m s$^{-1}$ difference in this layer.

As time progresses and the cores and gaps of precipitation begin to form at the equatorward end, $\partial v/\partial z$ (cross-frontal vertical shear) becomes more similar between the poleward and equatorward ends of the front (Fig. 7a). Meanwhile, $\partial u/\partial z$ (along-frontal vertical shear) at the poleward end remains nearly double that at the equatorward end ($40–45$ m s$^{-1}$ from the surface to 600 hPa at the poleward end vs $20–25$ m s$^{-1}$ at the equatorward end, Fig. 7b). Therefore, above the surface cold front, the along-frontal shear dominates throughout the simulation at the poleward end, indicating near-thermal wind balance, whereas cross-frontal shear becomes almost as great at the equatorward end.

To relate this contrast in the vertical wind profile to the precipitation distribution, the method of identifying the cold front through time described above also allows for an analysis of along-front variability of precipitation and other variables in the area just ahead of the front (where maximum precipitation falls, Fig. 3). The eastern bold line in Fig. 3 is obtained by moving 6 grid points

![FIG. 6. Hodographs at 21 h of the mean vertical wind profile along the front (the rear bold line in Fig. 3) at the poleward and equatorward ends separately. The marked values are the pressure levels (hPa).](image)

![FIG. 7. Time series of 600-hPa wind speed minus 10-m wind speed, showing the (left) $u$ and (right) $v$ components. Each value is calculated from taking the mean difference between 10-m and 600-hPa wind speed along the front (the rear bold line in Fig. 3) at the poleward and equatorward ends separately.](image)
(a) Precip

(b) 10-m vorticity

(c) 950-hPa vorticity

FIG. 8. Along-frontal variances of the given variables calculated every hour from 21 h onward between the two bold lines in Fig. 3. Variances are calculated along each of the six front-parallel lines between the two bold lines separately and then the mean of these six is plotted. Time series are calculated for the poleward and equatorward ends of the front separately.

(24 km) east of the identified front for each y coordinate. The space between the two lines is just wide enough to contain the cold-frontal precipitation, including later in the simulation when the precipitation cores form. Thus, analyzing along-frontal variability of precipitation and other atmospheric variables within these $24 \times 360 \text{ km}^2$ boxes for the poleward and equatorward ends of the front separately describes the variability of the precipitation cores and associated flow fields, both along the front and through time.

The time series of the variance of precipitation along the front captures when the precipitation cores start to become pronounced at the equatorward end after about 30 h (Fig. 8a). By contrast, the variance at the poleward end is near-constant through the simulation, reflecting the persistence of the continuous convective line of precipitation. The time series of 10-m vorticity variance along the front does not capture this rapid increase after 30 h (Fig. 8b). Furthermore, the 10-m vorticity variance at the poleward end is greater than at the equatorward end, despite the fact that the precipitation cores form only at the equatorward end. However, just above, rather than at the surface, the along-frontal vorticity variance at the equatorward end closely resembles the precipitation variance (Fig. 8c), with the precipitation and vorticity variances rapidly increasing at the same time and by the same proportional amount.

The greater dependence of equatorward precipitation on vorticity just above than at the surface is further illustrated by the fact that, after 30 h when the precipitation variance rapidly increases, precipitation is better correlated with both 950-hPa and even 900-hPa vorticity than with 10-m vorticity (Fig. 9b). Meanwhile, at the poleward end, the correlation of vorticity with precipitation steadily decreases with height (Fig. 9a). Therefore, precipitation at the equatorward end is aligned along the 950–900-hPa wind field, while precipitation at the poleward end is aligned along the surface wind field. The remainder of this paper investigates why the equatorward precipitation is less driven by surface winds and how the synoptic environment creates different vertical structures between these two parts of the front.

5. Synoptic differences between poleward and equatorward ends in lower-tropospheric winds and resulting differences in the vertical structure of a cold front

This section investigates what features of the flow environment, other than at the surface, determine the morphology of precipitation. As discussed in the introduction, the morphology of precipitation cores is commonly thought to depend largely on the magnitude and
orientation of near-surface vertical shear. Naturally, a front that is in near-thermal wind balance, with vertical shear oriented along it will be host to updrafts and hence precipitation remaining along the front. A front that has cross-front-oriented vertical shear will be host to updrafts that become tilted ahead of the front with height, so that some precipitation falls ahead of the front.

The wind and thermal fields between 950 and 850 hPa are markedly different between the poleward and equatorward ends of the front (Fig. 10). At the poleward end, there is a pronounced frontal structure and wind shift up to 850 hPa. At the equatorward end, there is no frontal structure or wind shift above 950 hPa. Consequently, the winds aloft are much more oriented across the surface cold front at the equatorward end. Consequently, updrafts remain erect at the poleward end and precipitation keeps falling uniformly along the front (Fig. 11, top panels). The layer of upright ascent (Fig. 11b) corresponds to the layer of almost purely meridional vertical shear at the poleward end (Fig. 6). Contrastingly, at the equatorward end, the updrafts become forward tilted by the cross-frontal winds aloft and the precipitation cores magnify (Fig. 11, bottom panels). By 35 h, when the precipitation cores are most pronounced, the hydrometeors are falling increasingly ahead of the surface cold front with height, so that maximum precipitation reaches the surface ahead of the front (Fig. 11d). The shallow depth of these circulations is consistent with previous studies of core-and-gap structures along cold fronts (e.g., James and Browning 1979; Hobbs and Biswas 1979; Hagen 1992; Browning and Reynolds 1994; Brown et al. 1999; Wakimoto and Bosart 2000; Jorgensen et al. 2003) and studies of continuous narrow cold-frontal rainbands (e.g., Browning and Pardoe 1973). However, the minimum contour of hydrometeors is 0.2 g kg$^{-1}$ to highlight those associated with the precipitation core, so precipitation particles are in fact falling throughout the lower troposphere.

Of course, this vertical structure at the equatorward end of the cold front only holds where the precipitation becomes displaced from the surface cold front. A strong contrast in the vertical frontal structure exists along the length of individual precipitation cores. There is no great contrast in the 2-m frontal temperature gradient along the length of precipitation cores (Fig. 12a) (i.e., whether frontal precipitation is heavy or light, the surface temperature gradient is roughly the same) and the surface frontal structure remains continuous along the length of precipitation cores and gaps (Fig. 13a).

At 900 hPa, on the other hand, the $\theta$ gradient is considerably greater where surface precipitation rate is greater (Fig. 12b), roughly 3 times greater where precipitation is $>5$ mm h$^{-1}$ (cores) than where precipitation is $<2$ mm h$^{-1}$ (gaps). This contrast in the thermal structure between cores and gaps is illustrated in Fig. 13 in the area indicated in Fig. 11c. At the poleward end of a typical precipitation core, the front and updraft are tilted eastward up to 800 hPa (Fig. 13b) as shown in Fig. 11d, so that the precipitation is farthest ahead of the surface cold front along the length of the core (Fig. 13a). Farther equatorward, where the precipitation is falling closer to the surface cold front, the frontal structure and updraft extend almost as high above the surface, but are erect (Fig. 13c), as with the continuous convective line along the poleward part of the front (Fig. 11b). In gap regions (i.e., where there is no heavy precipitation at or ahead of the surface cold front) the frontal structure and updraft are erect, but only extend to 900 hPa (Fig. 13d; note that hydrometeors are present, but below the minimum contour interval).

This contrast in the vertical structure of temperature and vertical velocity along the length of individual cores...
FIG. 10. Wind vectors (m s$^{-1}$) and $\theta$ (red contours every 2 K) at the given vertical levels in the poleward and equatorward boxes after 21 h. The top row shows 10-m winds and 2-m temperature.
indicates that the precipitation cores form in conjunction with a perturbation along the front, whereby surface frontogenesis is relatively uniform along the front, but occurs up to about 850 hPa where there are cores, and only slightly above the surface where there are gaps of precipitation. This perturbation does not occur at the poleward end of the front, due to the much deeper anastomosing cold-frontal structure with strong along-frontal wind shear keeping the updraft erect above the surface.

Despite the results of the simulations in the current study, the greater tendency for precipitation cores at the equatorward than poleward end is not universal. As stated in the introduction, Browning and Roberts (1996) documented a cold front with precipitation cores at the poleward but not equatorward end (their Fig. 19). Similarly, Kawashima (2016) performed real-data WRF simulations of a case where precipitation cores formed more distinctively and for longer at the poleward than equatorward end (his Fig. 12). Therefore, the equatorward part of the front is not necessarily where the most distinctive structures form. The particular poleward/equatorward contrast exhibited in this study is due to the particular baroclinic wave simulated and the particular synoptic environment to which this baroclinic wave leads. Crucially though, as with Browning and Roberts (1996), the precipitation cores in this study form along the part of the front where temperature and pressure contours are more perpendicular (Fig. 3), hence weaker thermal wind balance and a more kata-type front. As discussed previously, precipitation is more continuous along the front (i.e., less tendency for cores and gaps) where the front is closer to thermal wind balance, which may be at the poleward or equatorward end, depending on the synoptic setup as the cyclone develops.

6. Sensitivity of precipitation cores to thermal and frictional properties of the lower boundary

Although precipitation cores form at the equatorward end in CNTL, the surface temperature gradient does not vary in the along-frontal direction and precipitation remains continuous along the front (Fig. 11c), unlike in
some of the radar images in Fig. 1, where large gaps of absent precipitation form between cores. This section shows the effect of altering the heat and momentum fluxes from the lower boundary on the frontal flow environment and hence how some of the more distinctive precipitation structures may form along the front. As described in section 2, in the sensitivity simulations the front is forced to pass over a coastline, east of which either SST or roughness length is altered. Section 6a investigates the effect of altering SST and hence surface sensible heat fluxes, whereas section 6b investigates the effect of altering roughness length and hence surface momentum fluxes. The analysis is performed at the equatorward end of the front only, where the core-and-gap morphology is most pronounced in all simulations.

a. Sensitivity to SST

In PLUS2K, PLUS1K, MINUS1K, and MINUS2K, the SST of the land surface east of the coastline is increased or decreased by 1 or 2 K (as described by the name of each simulation) at each grid point. A few hours after the front passes over the coastline, the differences in the wind profile above the surface cold front are shown in Fig. 14a. Relative to CNTL, PLUS2K and MINUS2K slightly reduce and increase $du/dz$, respectively, up to about 700 hPa. The 1000–800-hPa $u$ shear in MINUS2K is about double that in PLUS2K. However, along-frontal winds are almost completely unaffected by SST (i.e., very little difference in $du/dz$ between simulations), so that in PLUS2K the along-frontal shear is relatively dominant. Consequently, for most of the simulations, greater SST implies maximum precipitation falling closer to the front (Fig. 15a). Toward the end of the simulations, those with lower SST show a decrease in the distance of maximum precipitation from the front on account of the fact that the precipitation rate along the front decays (not shown). Along-frontal variance of precipitation does not show that cores are more pronounced with lower SST (Fig. 15b), but variance is also affected by the fact that precipitation rate is greater when SST is greater.

The contrasts between PLUS2K, CNTL, and MINUS2K are illustrated in Fig. 16. Although the surface front is no different between the simulations, the less stable stratification with greater SST is shown by the $u$ contours (right panels), resulting in the greater ascent above the surface front, hence greater precipitation rate along the front (left panels). However, the lack of vertical shear of the zonal wind in PLUS2K (Fig. 14a) keeps precipitation in a continuous convective line, whereas cores and gaps form in the simulations with lower SST (Fig. 16, left panels). Therefore, enhanced sensible heat fluxes suppress the core-and-gap morphology.

b. Sensitivity to roughness length

In LANDFRIC1, LANDFRIC2, and LANDFRIC3, the roughness length of the land surface is progressively increased. The effect of greater friction on the vertical wind profile above the surface front is more striking than for SST variability (cf. Figs. 14a,b). The reduction of surface meridional wind speed with greater friction implies less meridional shear from the surface to the mid-troposphere, whereas the reduction of surface zonal wind speed implies greater zonal shear. Consequently, the shallow layer of meridional shear up to 900 hPa in CNTL is almost nonexistent in LANDFRIC2, in which

![Figure 12](url-to-figure)
zonal shear dominates the vertical wind profile from the surface to the midtroposphere. The effect in both alongfrontal variance of precipitation and distance of maximum precipitation from the front is striking (Figs. 15c,d). The time series for LANDFRIC2 is noisy because the line of maximum vorticity from which the location of the front is identified is not so well-defined in this simulation.

The greater alongfront variability of precipitation with increasing roughness length is illustrated in Fig. 17. As roughness length increases between simulations, the surface wind shift (left panels) and surface cold front (right panels) shift (Fig. 14).
panels) are increasingly poorly defined, as in the idealized simulations of Hines and Mechoso (1993), Rotunno et al. (1998), and Norris et al. (2014). In LANDFRIC3 there is hardly a surface wind shift at all and no discernable surface cold front (LANDFRIC3 was omitted from previous figures because the line of maximum vorticity from which the front is identified does not exist in this simulation). Therefore, a transition is evident between simulations in the cross sections: in CNTL the ascent is relatively erect above the surface cold front with hydrometeors falling relatively near the front; in LANDFRIC1, a separate maximum of vertical velocity has formed ahead of the front; in LANDFRIC2, this maximum ahead of the front has separated from the weak ascent above the front; in LANDFRIC3 almost all the ascent and hydrometeors are ahead of the front (the front is not even visible in the \( u \) contours). The more poorly defined cold front with increasing roughness length has also decreased static stability, enhancing updrafts, so that the precipitation rate increases from CNTL to LANDFRIC3.

LANDFRIC3 shows what can happen to the precipitation distribution along the cold front when the surface wind shift breaks down and mesovortices are allowed to dominate the surface wind field (Fig. 18). After 22 h, the strip of maximum 10-m vorticity representing the surface wind shift is still relatively intact and cold-frontal precipitation is continuous along it. Thereafter, the vorticity strip breaks up into the mesovortices and the core-and-gap morphology becomes increasingly distinctive. Unlike in CNTL, where the vorticity strip remains intact and a line of precipitation remains along the front, despite the formation of the cores (Fig. 5), large gaps of absent precipitation form between the cores in LANDFRIC3 (Fig. 18). The large precipitation cores in LANDFRIC3 resemble the long curved precipitation cores observed on radar in Figs. 1a and 1c, suggesting that these structures may have formed due to the cold front traveling a long distance over a rough land surface, with the surface wind shift and cold front breaking up into mesovortices.

These sensitivity simulations are consistent with the sensitivity simulations of Kawashima (2011). In simple idealized experiments, he found that greater vertical shear of the cross-frontal wind (\( du/dz \)), relative to the vertical shear of the along-frontal shear (\( dv/dz \)), enhances the growth rate and amplitude associated with cores and gaps. The current study has shown that the same holds in a more realistic primitive equation model and how in the real
In particular, lower surface heat fluxes may increase \( \frac{du}{dz} \) (Fig. 14a), whereas greater surface friction both increases \( \frac{du}{dz} \) and reduces \( \frac{dy}{dz} \) (Fig. 14b), so that the greatest differences between precipitation cores occur due to differences in roughness length.

7. Summary

Precipitation cores of anticyclonic orientation are frequently observed on radar along surface cold fronts of varying width, length, curvature, wavelength, and angle made with the surface cold front. Previous studies have related variability in the morphology of precipitation cores to variability in the wind and thermal profiles above the surface cold front. These studies have either been observational studies, in which the sensitivity of the cores to the atmospheric conditions is not investigated, or simple-idealized-model studies, in which the horizontal homogeneity of the large-scale deformation prevents an analysis of how the synoptic environment leads to differences in
precipitation morphology between different parts of the front.

In this study, a more realistic primitive equation model (WRF) was run at high resolution to investigate the sensitivity of cores in an idealized framework. Moist idealized baroclinic-wave simulations were performed, in which a mature cold front at 20-km grid spacing, subject to heat and momentum fluxes from the lower boundary, was reinitialized with a nested domain of 4-km grid spacing inserted to simulate

Fig. 17. As in Fig. 11, but comparing simulations of various roughness lengths.
FIG. 18. As in Fig. 5, but for LANDFRIC3.
clockwise-oriented precipitation cores along the cold front.

In the control simulation, where the lower boundary has roughness length appropriate for the ocean and a SST distribution equal to the initial temperature at the lowest model level, a continuous narrow cold-frontal rainband persists along the poleward part of the front, whereas at the equatorward end of the front periodic precipitation maxima appear within the narrow rainband, resembling those observed on radar.

The precipitation cores bear resemblance to counterpropagating mesovortices on either side of the front. The postfrontal mesovortices form on a line of vorticity behind the cold front, whereas the prefrontal mesovortices are attached to the line of maximum vorticity along the cold front. These mesovortices generate undulations in the lighter prefrontal and postfrontal precipitation and eventually interact across the cold front to generate the precipitation cores along the front. However, the precipitation cores are aligned along the winds just above rather than at the surface, indicating that ascent at the surface is unaffected, whereas ascent slightly aloft is rotated by the cross-frontal winds, tilting updrafts. These cross-frontal winds are associated with cross-frontal vertical shear, which is of similar magnitude at both the poleward and equatorward ends of the front. However, at the equatorward end, there is a weaker cold front and vertical shear of the along-frontal winds is hardly any larger than that of the cross-frontal winds. Contrastingly, at the poleward end, although the cross-frontal shear is similar to that at the equatorward end, the along-frontal shear is more than double the magnitude of the cross-frontal shear, associated with a more ana-type front that persists through the simulation and keeps updrafts erect.

At the surface, the line of maximum vorticity remains intact along both the poleward and equatorward parts of the surface cold front throughout the control simulation and, despite the formation of the precipitation cores at the equatorward end, precipitation remains continuous along the front. However, sensitivity simulations reveal that, when surface friction is greater and the surface wind shift across the front breaks down, the mesovortices dominate the surface wind field (as well as aloft), so that large gaps of absent precipitation form between cores. In simulations with the highest friction, the cold front eventually becomes poorly defined, so that there is no coherent structure to the vorticity or precipitation maxima. The sensitivity simulations also reveal that greater SST and hence reduced static stability, although not affecting the surface front, leads to more erect updrafts above the surface front and hence a more continuous rainband with no distinctive precipitation cores.

All the simulations in this paper illustrate that mesovortices may form, both ahead of and behind a cold front. When a cold front has a well-defined wind shift and temperature gradient extending well above the surface, the magnitude of vorticity in these mesovortices is much smaller than that along the cold front, so that the strip of vorticity along the cold front remains intact, updrafts remain erect, and precipitation keeps falling along the cold front. When the wind shift and temperature gradient rapidlly decay with height and there is little vertical structure to the cold front, mesovortices may dominate the wind field and hence vertical velocity just above the surface. However, as long as the surface cold front remains intact, the rainband remains continuous along the surface cold front, despite the formation of maxima within it, and the surface cold front provides a medium along which perturbations can propagate. In some cases, the wind shift and temperature gradient may weaken dramatically, including at the surface, allowing the mesovortices to dominate the full three-dimensional flow field, so that large gaps of precipitation appear along the front. In this case, there is no surface cold front for the cores to propagate along and the precipitation field becomes increasingly poorly defined. This paper has shown that this case may occur in the event of high friction, greatly reducing surface winds and leading to a poorly defined cold front.

These sensitivity simulations may explain much of the observed variability in the core-and-gap morphology along cold fronts. As shown on radar and by these simulations, the morphology of precipitation cores may vary along the cold front at a given time, over time during a cold front’s evolution, and between different cold fronts. These simulations demonstrate how the precipitation cores depend on variations in the synoptic environment, and the resulting variations in the wind and temperature profiles above the surface cold front.

Acknowledgments. 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 was a NERC-funded student through the Diabatic Influences on Mesoscale Structures in Extra-tropical Storms (DIAMET) project, NE/1005234/1. Vaughan and Schultz were partially funded by NERC through DIAMET. We thank the anonymous reviewers for their comments that improved this article.
REFERENCES


