To understand near-surface strong winds in extratropical cyclones, a simulation of a dry, idealised baroclinic wave is presented. The forces that accelerate the winds are analysed using the terms in the horizontal momentum equation. Two regions of strong near-surface winds developed within the simulation: one to the east of the low centre and the other to the southwest. The flow to the east, resembling the cold conveyor belt, accelerated as it passed through the strong pressure gradient associated with the warm-frontal zone. This acceleration was reduced by friction near the surface. The winds to the southwest were characterised by three airstreams. One airstream consisted of air parcels that started north of the warm front near the surface, accelerated north of the warm front, encircled the low and continued to accelerate to the southwest of the low. The increases in wind speed in this airstream (also part of the cold conveyor belt) resulted from the along-flow pressure gradient force being greater than friction. The second airstream consisted of air parcels that descended west of the cold front and maintained their speed during descent until they were slowed either by friction in the boundary layer, or as they moved southward against the local pressure gradient. Between these two airstreams, a third airstream was characterised by air parcels that accelerated during descent from the mid to lower troposphere whilst moving around the bent-back front. This airstream bears some resemblance to the sting jet and was accelerated by the pressure gradient force on descent until it encountered increased friction or moved against the local pressure gradient.

Key Words: windstorm; extratropical cyclone; sting jet; cold conveyor belt; rapid deepening; momentum equation

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1. Introduction

Strong windstorms in Europe can cause large swathes of damage. For example, Windstorm Klaus, which passed over France and northern parts of Spain in 2009, and Windstorm Xynthia, which crossed over much of Western Europe in 2010, each caused insured losses in excess of 1 billion euros (Zuba, 2012). Furthermore, Xynthia killed 65 people across Europe – many as a result of the associated storm surge in coastal areas (Kolen et al., 2010). Windstorm Gudrun deprived 700 000 people in Sweden of electricity in January 2005, some 56 000 of whom did not have their electricity restored for 8–20 days (Sundell et al., 2006). Fifty-three percent of all damage to European forests caused by natural disturbances has been attributed to storm damage, mostly from large windstorms, an estimated 18 million m$^3$ of wood annually (Schelhaas et al., 2003).

Because of the high societal and economic impact of intense extratropical cyclones, they have formed the focus of numerous studies in the literature. In an investigation of the 1992 New Year’s Day storm that made landfall on the west coast of Norway, Gronås (1995) found the strongest winds occurred to the south of the low centre associated with a bent-back front. The location of the strongest winds to the south of the low was also found in a three-year study of hurricane-force winds in extratropical cyclones by Von Ahn et al. (2005).

Studying the Great Storm of 15–16 October 1987, Browning (2004) found that the strongest surface winds occurred to the south of the low and east of a region of cloud banding at the tip of the cloud head, a phenomenon he termed a sting jet. Observations of this banding and associated mesoscale slantwise circulations led Browning (2004) to suggest that wind maxima to the south of the low could be explained by evaporative cooling or conditional symmetric instability (CSI; e.g. Bennetts and Hoskins, 1979; Emanuel, 1983; Schultz and Schumacher, 1999). Wind maxima in strong extratropical cyclones have also been attributed to the release of CSI by Gray et al. (2011).

However, when Smart and Browning (2014) simulated a strong extratropical cyclone that crossed the UK on 3 January...
2012, they found that ‘the simulations did not support the idea that either the evaporation or CSI played a major role’ in the development of the wind maxima. These contrasting results suggest the need to better understand causes of damaging winds within extratropical cyclones.

In this article, we examine the wind field in an idealised extratropical cyclone. Extratropical cyclones can be studied in a simplified framework using an idealised baroclinic wave. Here, an unstable initial state is perturbed, and integrated forward in time using the primitive equations, resulting in an extratropical cyclone (e.g. Mudrick, 1974; Simmons and Hoskins, 1978; Takayabu, 1986). The advantage of this technique is that realistic features develop from an analytically prescribed state, so a hierarchy of complexity of physical processes can be investigated. Baker et al. (2014) examined the wind field in an idealised framework, focusing on the role of CSI in strengthening winds within a cyclone. However, CSI requires the presence of moisture within the cyclone. Here, we examine the wind field of a cyclone in the absence of moisture and therefore the absence of moist instabilities. We pose the question: what is the wind field produced by a dry, idealised baroclinic wave?

Cold and warm conveyor belts (Carlson, 1980) have been produced in dry, idealised baroclinic waves (Golding, 1984). Indeed, dry, idealised baroclinic waves develop wind maxima both southwest and east of the low (Takayabu, 1986; Polavarapu and Peltier, 1990). But what mesoscale wind features can be simulated in a dry, idealised extratropical cyclone? And how do these wind features develop?

Papritz and Schemm (2013) highlighted the role of vertical ageostrophic geopotential fluxes in the development of a low-level jet in a dry, idealised downstream cyclone. However, previous studies do not examine the forces that affect the wind speed directly. To diagnose the regions of strong winds in extratropical cyclones, we use the horizontal momentum equation, which relates the acceleration of an air parcel to the forces acting on it. The magnitude and direction of these forces can be obtained directly from numerical weather prediction model output. Previously, this technique has been applied to cold-air damming (Bell and Bosart, 1988), an Australian southerly buster cold front (McBride and McInnes, 1993), cold-air surges (Coffe and Mass, 1995), gap flow over the Gulf of Tehuantepec (Steinburgh et al., 1998), an easterly tip jet near Greenland (Outten et al., 2009), and a mesoscale convective system (Mahoney et al., 2009).

In this article, we use the horizontal momentum equation as the principal diagnostic framework to analyse the forces causing regions of strong winds in a dry, idealised extratropical cyclone. This article begins with a description of the model setup and initial condition in section 2. The momentum equation diagnostics that have been developed are then introduced in section 3. A brief overview of the development of the cyclone is given in section 4, with a particular focus on the evolution of regions of strong winds around the cyclone. The space and time development of the forces responsible for local accelerations of strong winds are examined in section 5. To understand the accelerations which air parcels experience, a Lagrangian description using air-parcel trajectories is then introduced in section 6. The role of frontalities in causing the descent of some of these trajectories is briefly discussed in section 7. Our results are compared to conceptual models and previous observations in section 8, and the article is summarised in section 9.

2. Model set-up

This study employs the Advanced Research Weather and Forecasting Model Version 3.4 (WRF-ARW; Skamarock et al., 2008) in an idealised, baroclinic, periodic, channel set-up. The simulation is performed on an f-plane, neglecting spherical geometry, topography and moist processes. The model domain has a grid spacing of 20 km over an extent of 4000 km zonally and 8000 km meridionally. There are 40 vertical levels. The grid spacing in this study is coarser than that used in previous studies of strong winds (e.g. Martínez-Alvarado et al., 2010; Smart and Browning, 2014; Baker et al., 2014). To test the sensitivity of our simulation to the horizontal and vertical grid spacing, the simulation is also run at a 20 km grid spacing with 80 levels, a 10 km grid spacing with 40 levels and a 10 km grid spacing with 80 levels. The resulting wind fields are nearly unchanged, so we choose the 20 km grid spacing because this spacing is sufficient to resolve mesoscale circulations, but without producing noisy fine-scale flow structure that would complicate our analysis of the relevant physical processes. At the lower boundary, a time-invariant, meridionally varying temperature field is specified using a pressure-weighted linear extrapolation from the lowest three model levels of the initial condition.

The initial zonal jet is in thermal-wind balance and is obtained by inverting a potential vorticity (PV) field specified in the y–z plane (e.g. Heckley and Hoskins, 1982; Davis and Emanuel, 1991; Rotunno et al., 1994). Our PV field is similar to that used by Plougonven and Snyder (2007), except we specify a uniform value of 0.4 PVU in the troposphere and 5 PVU in the stratosphere to strengthen the jet. The initial jet is then extended in the x-direction to give a three-dimensional initial condition that has a maximum zonal upper-level jet speed of 75 m s⁻¹. To this state, a potential temperature perturbation θ' is added (e.g. Polvani et al., 2004; Schultz and Zhang, 2007). This perturbation takes the form of

$$\theta'(x, y, z) = A \cos \left( \frac{2\pi x}{L_x} + \pi \right) \cos^2 \left( \frac{\pi y}{L_y} \right),$$

where A is the amplitude of the perturbation (taken to be 1 K) and \(L_x = 4000\ km\). The radius of the perturbation is given by

$$r = \left( \frac{y - z_0}{r_y} \right)^2 + \left( \frac{z - z_0}{r_z} \right)^2 \frac{1}{2},$$

where \(L_y = 8000\ km\), \(z_0 = 8\ km\), \(r_y = 2000\ km\), and \(r_z = 8\ km\). The model employs the MMS similarity surface physics (Skamarock et al., 2008) with an ocean surface. The YSU planetary boundary-layer scheme (Hong et al., 2006) is used with surface fluxes of momentum, heat and moisture. As our goal is to study the structure and evolution of the wind field in a dry, idealised baroclinic wave, the model was run without a microphysics scheme, a convective scheme or an explicit diffusion scheme. Furthermore, relative humidity was set to zero as an initial condition and the effect of moist surface fluxes was reduced by turning latent heat release off at run-time, so the only effect of moisture in our simulation is the perturbation of density by moist surface fluxes.

A baroclinic wave without fluxes of moisture and heat was also simulated. In this no-fluxes simulation, the low-level, southerly warm-sector air was not cooled whilst it flowed northward over the colder ocean. Similarly, the low-level, northerly flow was not warmed whilst it moved over the warmer ocean southwest of the low. This lack of low-level heating and cooling led to the static stability being artificially increased southwest of the low and reduced in the warm sector. The resulting baroclinic wave developed weaker fronts and a weaker, less realistic cyclone than the simulation with latent heat release turned off and fluxes of moisture and heat, so the simulation presented here includes surface fluxes.
3. Momentum equation diagnostics

The horizontal momentum equation is formulated as

$$\frac{\partial u_h}{\partial t} = - (u_h \cdot \nabla) u_h - \frac{w}{\rho} \frac{\partial u_h}{\partial z} \tag{3}$$

where $u_h = (u, v)$ is the horizontal wind, $\nabla_h = (\partial/\partial x + j \partial/\partial y)$ is the horizontal gradient operator, $\partial u_h/\partial t$ is the local acceleration of the horizontal wind, $w$ is the vertical velocity, $f$ is the Coriolis parameter, $\rho = \rho(x, y, z)$ is the density, and $p = p(x, y, z)$ is the pressure. There are six terms on the right-hand side: HADV is the horizontal advection term of horizontal momentum, VADV is the vertical advection of horizontal momentum, $F_{\text{COR}}$ is the Coriolis force, $F_{\text{PGF}}$ is the pressure gradient force, $F_{\text{PBL}}$ is a term produced by the boundary-layer parameterization scheme, and $RES$ is a residual term. The boundary-layer term $F_{\text{PBL}}$ is calculated by taking the momentum tendency from the boundary-layer scheme, and so includes the effects of friction and vertical diffusion. The local acceleration $\partial u_h/\partial t$ is calculated from hourly model output fields using a centred-difference method to allow the computation of the residual term $RES$. This residual includes errors arising from the time-interpolation of $\partial u_h/\partial t$ and implicit horizontal diffusion. (The model is run without explicit horizontal diffusion.) All other terms on the right-hand side of Eq. (3) (HADV, VADV, $F_{\text{COR}}$, $F_{\text{PGF}}$ and $F_{\text{PBL}}$) are output directly from the model.

The terms in the momentum equation are calculated from the model to determine how the strong winds are accelerated. On any given horizontal level, the only terms leading to an increase in the maximum wind speed are vertical advection VADV and the pressure gradient force $F_{\text{PGF}}$. The horizontal-advection term HADV redistributes momentum over the horizontal plane, the Coriolis force $F_{\text{COR}}$ acts only to turn the winds, and friction from the boundary layer $F_{\text{PBL}}$ retards the flow. In this article, we focus on the relative contributions of vertical advection, the pressure gradient force and friction to the along-flow acceleration and hence changes in wind speed. These forces can also be analysed in a Lagrangian framework with the momentum equation cast as

$$\frac{d\mathbf{u}_h}{dt} = - \mathbf{f} \times \mathbf{u}_h - \frac{1}{\rho} \nabla_h p + F_{\text{PBL}} + RES, \tag{4}$$

where the left-hand side represents the material acceleration of an air parcel. Air-parcel trajectories are calculated offline using the WRF Software Read/Interpolate/Plot (RIP; Stoelinga, 2009). The trajectories are determined from model-derived velocity fields that are output every hour. The accuracy of the trajectories is improved by using a 10 min time step in RIP (Stoelinga, 2009). Variables along the trajectory are interpolated from the hourly model fields to the air-parcel location at each RIP time step. Bilinear interpolation is used horizontally, whereas linear interpolation is used vertically and in time. There are no advection terms in the Lagrangian framework, so the only along-flow accelerations in Eq. (4) are provided by the pressure gradient force and friction.

4. Overview of cyclone evolution

From the initial state, a thermal wave developed (Figure 1). By Day 4.75, the low had deepened to 986 hPa (Figure 1(a)). A warm front developed east of the low and a bent-back front extended to the west of the low centre (Shapiro and Keyser, 1990; Schultz and Vaughan, 2011). The developing cold front, south of the low, was noticeably weaker than the warm front, a feature thought to be representative of rapidly deepening marine cyclones (Hines and Mechoso, 1993). Frontal fracturing occurred by Day 5.23 as the low deepened to a central pressure of 972 hPa (Figure 1(b)). By Day 5.75, the low centre was well separated from the 'T-bone' between the warm and cold fronts, which both intensified through to Day 6.25 (Figure 1(c,d)). After Day 6.25, the cyclone continued to wrap up and intensify, extending the length of the occluded front.

At Day 4.75, a wind-speed maximum of over 20 m s$^{-1}$ occurred to the southwest of the low (Figure 1(a)). A second wind-speed maximum of over 15 m s$^{-1}$ appeared northeast of the warm front. This easterly wind-speed maximum strengthened to over 20 m s$^{-1}$ by Day 5.25 and was confined beneath the warm-frontal zone (Figures 1(b) and 2(a)). At Day 5.25, the strongest winds remained southwest of the low and were connected to the strong winds at upper levels (Figures 1(b) and 2(a)).

As the cyclone developed, the wind-speed southwest of the low reached over 35 m s$^{-1}$ by Day 5.75, and the wind speed to the east of the low increased to over 30 m s$^{-1}$ (Figure 1(c)). Further west, a maximum in wind speed of 30 m s$^{-1}$ appeared around the bent-back front centred at 875 hPa (Figure 2(b)). This feature was embedded in a region of winds greater than 20 m s$^{-1}$ that extended from the surface up to the upper-level jet. At Day 6.25, the strongest winds remained to the southwest of the low with a maximum value of 38 m s$^{-1}$ (Figure 1(d)).

To first order, the winds respond to the pressure gradient, so the stronger the pressure gradient, the stronger the wind. Figure 3 shows that this relationship is largely true. Closer examination reveals a more complicated relationship between central pressure, maximum pressure gradient, and the maximum horizontal wind speed. The peaks in horizontal wind speed at Day 5.5 and Day 6.1 lagged the peaks in the pressure gradient force. Furthermore, the peaks in the maximum horizontal pressure gradient were sharper than those in the horizontal wind speed. The blunter peaks in the maximum wind speed are a result of the wind speed being dependent on the integral of the pressure gradient over time, which acts to smooth out small-scale variations of the pressure field.

From Day 4.75 to Day 5.75 the cyclone deepened from 986 to 962 hPa (24 hPa (24 h$^{-1}$)). This deepening rate is less than in some observed and simulated strong cyclones (e.g. Gronás, 1993; Browning, 2004; Martínez-Alvarado et al., 2010; Smart and Browning, 2014), however the cyclone fulfilled the definition of a 'bomb' (Sanders and Gyakum, 1980) and developed regions of strong winds comparable to those in these other studies, so this idealised simulation provides a suitable example for our scientific problem.

5. The Eulerian perspective

Equation (3) can be used to investigate the relationship between the pressure gradient force and the horizontal wind speed in more detail. This equation gives the terms responsible for the local acceleration at each grid point. The along-flow forces (responsible for local increases in wind speed) are obtained by projecting these terms onto the unit vector of the horizontal wind (Figure 4). North of the warm front the pressure gradient force strengthened the easterly flow at Day 4.75 (Figure 4(a)). Along-flow positive local acceleration in the strong-wind region southwest of the low at Day 4.75 was provided both by vertical advection and the pressure gradient force (Figure 4(a)). At Day 5.25, the flow north of the warm front continued to be strengthened by the pressure gradient force (Figure 4(b)). The acceleration was greater nearer the warm front because the horizontal pressure gradient was maximum (Figure 1(b)). Northwest of the low, the flow ascended and weakened whilst being advected to the northwest of the low through a region where there was no positive along-flow pressure gradient (Figures 4(b) and also 13(a) below). Also, in
this region to the northwest of the low, the frontogenesis field changed from being frontogenetic to frontolytic (Figure 13(a) below). The ascent of the flow on the cold side of the bent-back front is consistent with an indirect circulation as the flow moved into the region of frontolysis (Schultz and Sienkiewicz, 2013). As the flow continued southwest and south of the low, the pressure gradient force increased the speed of the flow in a region co-located with the wind-speed maximum (Figure 4(b)). In a vertical cross-section, the along-flow pressure gradient force was strongest near the surface both north of the warm front and southwest of the low (Figure 2(a)). The region of along-flow vertical advection south of the low extended to 700 hPa with a maximum below the upper-level front (Figure 2(a, b)).

The area where the wind speed was being increased by the pressure gradient force to the low grew by Day 5.75 (Figure 4(c)). At this time, the speed of the flow was also increased by vertical advection in pockets to the southwest and south of the low centre. The along-flow vertical-advection term was large above the warm front as momentum from lower levels was carried up the warm-frontal surface (Figure 2(b)). The vertical-advection term also remained large beneath the upper-level front (Figure 2(a, b)).

The along-flow residual term at the time of the strongest winds (Day 6.25) at the lowest model level is contoured in Figure 5. This term is comparable with the acceleration terms only in two narrow strips well away from the regions of strong winds. The absence of a large residual in the southwest wind maximum suggests the error arising from implicit horizontal diffusion and time-interpolation of $\partial u_h/\partial t$ is small in this region. Furthermore, the relative size of the residual term, compared to the other five terms on the right-hand side of Eq. (3), decreased with height (not shown). As our focus here is on the acceleration of the strongest winds, we conclude that this acceleration is well represented by the explicit terms on the right-hand side of Eq. (3).

6. The Lagrangian perspective

An Eulerian perspective, given by Eq. (3), is useful in determining the relative contribution of vertical motion and the pressure gradient acceleration to the wind speed at a point in the cyclone. However, a Lagrangian perspective is required to understand the forces that individual air parcels experience (Eq. (4)). Air parcels were released at every grid point in the centre of the domain between 0 and 4000 km in the zonal direction and 3000 and 5000 km in the meridional direction (Figure 6). These air parcels were released at Day 6.25 and trajectories run backwards for 30 h to Day 5. We chose to release the air parcels at Day 6.25 because this was the time of maximum horizontal surface wind speed (Figure 3). Backward trajectory analysis was also carried out for Days 5.25 and 5.75 with similar results although the magnitude of the accelerations and descent the air parcels experienced were smaller than for Day 6.25. The minimum pressure that the air parcels attained within these 30 h was determined and subtracted from the pressure of the air parcel at Day 6.25 to give a measure of the descent within the cyclone during this period. A value of zero means that the minimum pressure was attained at Day 6.25, meaning that the parcel either ascended to a lower pressure surface or stayed on a constant pressure surface.

A horizontal map of this quantity shows a value of zero along a comma-shaped region encompassing the low centre and the warm sector, indicating ascent of air parcels (Figure 6). Descent

Figure 1. Pressure (bold contours every 4 hPa), potential temperature (thin contours every 5 K) and horizontal wind speed (m s$^{-1}$, coloured shading above 10 m s$^{-1}$) at lowest model level $\sigma = 0.972$ ($\approx 200$ m) for (a) Day 4.75, (b) Day 5.25, (c) Day 5.75, and (d) Day 6.25. Solid lines AB and CD represent locations of cross-sections in Figure 2.
and was co-located with the region south of the low centre where vertical advection provided a positive along-flow acceleration to the horizontal wind (Figure 4(d)).

6.1. Eastern wind maximum cross-section

An east–west vertical cross-section along the line EF in Figure 6 was constructed over the wind-speed maximum to the east of the low (Figure 7). A region of strong descent occurred east of the warm front. West of the warm front, a group of trajectories stayed level or rose over the warm front and entered the region of strong winds centred at 875 hPa. A subset of the trajectories were analysed to examine the behaviour of parcels entering the wind-speed maximum to the east of the low. This subset consisted of 45 trajectories taken from the first five model levels (with approximate altitudes 200, 600, 900, 1300, 1700 m) giving nine trajectories at each level. A representative subset of nine of these 45 trajectories is shown in Figure 8. The subset of 45 trajectories breaks into two groups.

The first group of trajectories includes A2–A9, B3–B9, C5–C9, and D9 (circles in Figure 7) and is represented by A2, A5, A9, B5, B9, and D9 in Figure 8. These trajectories entered the wind-speed maximum to the east of the low, approaching anticyclonically from between 1500 and 3000 km east of the low centre (Figure 8). They descended from the anticyclone to the northeast of the low centre, then rose slightly and accelerated quite substantially from between 5 and 10 m s\(^{-1}\) to between 30 and 40 m s\(^{-1}\) in 12 h (Figures 8 and 9(a,b)). This acceleration resulted from the stronger pressure gradient as the parcels approached the warm-frontal zone (Figures 8 and 9(e)). The pressure gradient force was large enough in the along-flow direction to overcome any friction (Figure 9(c,d,f)). The trajectories with larger wind speeds at Day 6.25 on any horizontal level finished closer to the cyclone centre and started farther south than trajectories that finished with lower wind speeds (e.g. A2 versus A5, A5 versus A9, B2 versus B9; Figures 8 and 9(a)).

The second group of trajectories includes A1, B1–B2, C1–C4, D1–D8, and E1–E9 (squares in Figure 7) and is represented by B2, D2, and D5 in Figure 8. These trajectories started farther south than the first group and ended within the wind-speed maximum above the warm front. The trajectories underwent relatively little vertical motion until they approached the cyclone and ascended anywhere from 70 to 130 hPa between Days 5.75 and 6.25 (Figures 8 and 9(b)). Like those in the first group, these trajectories experienced an increasing magnitude of the pressure gradient force, which led to rapid acceleration of between 20 and 30 m s\(^{-1}\) in 12 h (Figure 9(a,c,f)).

The increase in positive along-flow pressure gradient was more abrupt for some trajectories (e.g. A9, B9) than others (e.g. A2, B2) because of the angle of approach to the warm-frontal pressure trough (Figures 8 and 9(c)). For example, A2 and B2 experienced a gradually increasing along-flow pressure gradient force as the parcels travelled almost parallel to the warm front between Days 5.5 and 6.25, whereas A9 and B9 moved towards the warm front more orthogonally, experiencing a more rapid change in direction and an abrupt change in along-flow acceleration at Day 6 (Figure 9(a,c)).

Thus, the strong lower-tropospheric winds to the east of the low centre resulted from the increasing horizontal pressure gradient experienced by parcels as they left the predecessor anticyclone and approached the strong pressure gradient associated with the deepening low centre. For parcels near the Earth’s surface, friction somewhat offset the acceleration due to the pressure gradient force.

6.2. Southwestern wind maximum cross-section

A northwest–southeast vertical cross-section along the line GH in Figure 6 was constructed over the wind-speed maximum to

Figure 2. Vertical cross-sections through cyclone as shown in Figure 1. Vertical axis is pressure in hPa. Potential temperature (thin contours every 2 K), horizontal wind speed (m s\(^{-1}\), shaded above 20 m s\(^{-1}\)) for (a) Day 5.25 and (b) Day 5.75. The along-flow acceleration components are contoured at 1 m s\(^{-1}\) h\(^{-1}\) (dashed) and 4 m s\(^{-1}\) h\(^{-1}\) (solid), with blue representing the pressure gradient force, and red vertical advection. Arrows denote regions of ascent and descent. All fields are smoothed with a four-point horizontal window and a two-point vertical window.

Figure 3. Evolution of minimum pressure (hPa, MIN_P, solid), maximum horizontal wind speed (m s\(^{-1}\), MAX_HWIND, dotted) and maximum horizontal pressure gradient force (m s\(^{-1}\) h\(^{-1}\), MAX_PGF, dashed) at \(\sigma = 0.972 \approx 200 \text{ m} \).
the southwest of the low at Day 6.25 (Figure 10). The strongest near-surface winds of over 35 m s\(^{-1}\) occurred below 850 hPa in association with vertical isentropes indicative of a well-mixed boundary layer.

However, the strongest descent occurred to the southeast of the low in association with the cold-frontal zone. This region of descending mid-tropospheric air extended to the warm side of the front above 780 hPa and to the cold side of the front below 840 hPa. Two other maxima in descent were embedded in the region of strong winds around the bent-back front, one centred at 875 hPa and the other at 775 hPa (Figure 10).

A subset of the trajectories released from this cross-section, covering the region of strong winds, was analysed to examine the behaviour of parcels entering the wind-speed maximum. This subset consisted of 110 trajectories taken from the first five model levels, giving 22 trajectories at each level. A1–A10, B1–B10, C1–C7, D1–D6, and E1–E7 (squares in Figure 10) did not enter the southwest wind maximum and are not discussed here. The remaining trajectories break into two distinct groups and a third group representing the transition between these two groups.

One group of trajectories that entered the southwest wind maximum descended from the west of the low (e.g. A21–A22, B21–B22, C21–C22, D21–D22, and E20–E22; triangles in Figure 10) and are represented by A22, B22, and D22 in Figure 11. Between Days 5 and 5.5 these trajectories descended anywhere from 50 to 150 hPa, exhibiting a range of behaviours (Figure 12(a)). The higher trajectories (e.g. D22) accelerated as they encountered a stronger pressure gradient force whilst descending, and the lower trajectories (e.g. A22) slowed due to friction (Figure 12(a–d)). However, after Day 5.5, all trajectories in this group slowed between 1 and 21 m s\(^{-1}\) (Figure 12(a)). The higher trajectories slowed because they moved southward against the local pressure gradient (e.g. D22), and the lower trajectories slowed due to friction despite a positive along-flow acceleration from the pressure gradient force (e.g. A22) (Figure 12(c, d, f)).

The second group of trajectories that entered the southwest wind-speed maximum at Day 6.25 (A11–A18, B10–B18, C8–C18, D7–D18, and E8–E18; circles in Figure 10) are represented by A12, B12, and D12 in Figure 11. Similar to...
Figure 6. Horizontal wind speed (m s\(^{-1}\), grey shading above 10 m s\(^{-1}\)) shown for Day 6.25 at \(\sigma = 0.972\) (≈ 200 m). Blue contours (at \(-1, -25, -50, -100, -200\) and \(-300\) hPa) show the difference between the minimum pressure that a trajectory released backwards from a grid point at \(\sigma = 0.917\) (≈ 600 m) reaches over 30 h and that trajectory’s final pressure at Day 6.25. The \(-1\) hPa contour is used instead of the 0 hPa contour because the latter was noisy due to machine error in calculating the descent of the trajectories. A value of zero means that the parcel attained its minimum pressure at Day 6.25 and so either ascended to this final pressure or remained at this pressure for the trajectory’s 30 h duration. The low pressure centre at Day 6.25 is denoted by ‘L’. EF and GH in red denote the positions of cross-sections shown in Figures 7 and 10. All fields are smoothed with a four-point window.

Figure 7. Horizontal wind speed (grey shading above 20 m s\(^{-1}\)) and potential temperature (thin black contours every 2 K) are shown at Day 6.25 for a vertical cross-section along EF in Figure 6. Horizontal wind speed and potential temperature are smoothed with a four-point horizontal window and a two-point vertical window. Bold blue contours show trajectory pressure differences as described in Figure 6, but here at \(-25, -50\) and \(-200\) hPa with quintic smoothing applied to a \([nx, nz] = [150, 10]\) grid. The release positions of backward trajectories are numbered and shown with coloured symbols (group 1 – circles, group 2 – squares).

The trajectories that entered the eastern wind-speed maximum, these trajectories initially slowed as they moved away from the preceding anticyclone (Figures 11 and 12(a)). The parcels were then accelerated along the warm/occluded front by the pressure gradient force, increasing in speed between 20 and 30 m s\(^{-1}\) in 6 h (Figure 12(c, f)). Soon after Day 5.85, the trajectories moved to the north and northwest of the low and experienced a decrease in pressure gradient force (Figures 11 and 12(c)). The trajectories ascended between 10 and 120 hPa, consistent with the change in sign of the frontogenesis field to the northwest of the low (Figure 13(a)). The resulting decrease in along-flow pressure gradient force led to decreased speed as friction slowed the trajectories and, in the case of the higher trajectories (e.g. D12), the parcels ascended against the local pressure gradient (Figure 12(a, c, d, f)). After Day 6, the parcels moved around the bent-back front, the pressure gradient force increased, and the parcels accelerated between 10 and 20 m s\(^{-1}\) in 6 h (Figures 11 and 12(a, e)). This acceleration ended due to an increase in the along-flow friction near Day 6.25 (Figure 12(d)). These cyclonic trajectories (e.g. A12, B12 and D12) experienced a larger along-flow friction at Day 6.25 than other trajectories entering the southwest wind maximum (e.g. A20, B20, D20, A22, B22 and D22). This larger friction resulted from both a larger speed southwest of the low and a stronger thermal contrast between the parcels’ temperature and the ocean surface, consistent with a well-mixed boundary layer due to cold-air advection, and thus an increased turbulent momentum flux and reduced parcel speed (Figure 12(c)). The stronger thermal contrast was a consequence of...
of the parcels’ lower initial altitude and more northerly starting position than the other trajectories entering the southwest wind maximum (Figures 11 and 12(b)).

A group of transitional trajectories between those that entered the southwest wind maximum from west of the low and those trajectories that entered the southwest wind maximum cyclonically from the east existed (A19–A20, B19–B20, C19–C20, D19–D20, and E19; diamonds in Figure 10), represented by A20, B20, and D20 in Figure 11. These parcels descended into the southwest wind-speed maximum like the trajectories in the first group (Figure 12(b)). However, the air parcels were accelerated whilst taking a cyclonic path around the low like trajectories in the second group, so we call them transitional trajectories (Figures 11, 12(a)). These parcels started north of the low and ascended around the northwest of the low (Figures 11, 12(b)). The parcels slowed between 7 and 8 m s$^{-1}$ as they moved against the local pressure gradient force until Day 5.25 (Figure 12(a, c)). As the trajectories continued around the bent-back front, they accelerated and continued to accelerate whilst descending anywhere from 70 to 200 hPa into the southwest wind-speed maximum with speeds of between 29 and 34 m s$^{-1}$ (Figures 11 and 12(a, b)). For the lower trajectories (e.g. A20, B20), the acceleration ended when friction became large enough to oppose the positive along-flow pressure gradient force (Figure 12(c, d, f)). For the higher trajectories (e.g. D20) the period of acceleration ended both as a result of friction and southward movement against the local pressure gradient (Figure 12(c, d, f)).

In summary, the strong winds to the southwest of the low centre consisted of two primary airstreams and a third, transitional airstream. The first airstream brought strong winds from the mid-troposphere towards the surface, where parcels were slowed by friction or movement against the local pressure gradient as they turned south away from the low centre. The second airstream consisted of trajectories that approached the low from the east and were accelerated around the low into the southwest wind maximum. Parcels in this airstream were accelerated by the pressure gradient force that they experienced north of the warm front and around the bent-back front. The transitional airstream consisted of parcels that rose around the bent-back front and were then accelerated into the southwest wind maximum by the increasing pressure gradient force as they continued around the bent-back front. The acceleration of these parcels ended due to increased friction or movement against the local pressure gradient.

7. Frontolysis

Two of the trajectory groups that entered into the southwest wind-speed maximum descended, bringing strong winds near to the surface from aloft. Schultz and Sienkiewicz (2013) suggested that the descent of high-momentum trajectories could be forced by frontolysis near the bent-back front in Shapiro–Keyser cyclones. In the simulation presented here, the southwest wind-speed maximum initially developed southeast of a region of frontolysis associated with the bent-back front (Figure 13(a)) and beneath another region of frontolysis located near the upper-level jet (Figure 13(c)). However, at the time of maximum surface horizontal winds, frontolysis along the bent-back front had weakened and the wind-speed maximum was co-located with a region of weak frontogenesis (Figure 13(b)). This weakening of the region of frontolysis upstream of the wind maximum at the time of strongest surface wind speeds suggests that the forcing for the descent of high-momentum trajectories occurred prior to the time of strongest surface winds. Although the trajectories entering the frontogenetic part of the wind maximum (e.g. A12, B12, D12) did not undergo large descent, the trajectories that descended near to the surface from higher levels (e.g. A20, A22, B20, B22, D20, D22) entered the region of weak frontolysis behind the cold front (Figures 10 and 13(d)). Both the frontogenetic and frontolytic regions were co-located with a layer of close to neutral static stability (Figure 10). However, descending air parcels were only evident in the region of near-surface frontolysis, supporting the idea that frontolysis is a useful diagnostic in determining where strong winds are brought to the surface from aloft (Schultz and Sienkiewicz, 2013).
8. Strong winds in extratropical cyclones

Both the trajectories entering the eastern wind-speed maximum and the second group of trajectories entering the southwest wind-speed maximum started in the cold air north of the warm front, east of the low and remained below 800 hPa (Figures 8, 9(b), 11, and 12(b)). These trajectories resemble the cold conveyor belt described by Carlson (1980) that originates in the anticyclonic low-level flow to the rear of the surface high east of the surface low. The trajectories remained underneath the warm-frontal zone, approaching the low from the east and encircling the low centre. Furthermore, some of the trajectories in the second group rose rapidly as they approached the bent-back front. However, in contrast to the cold conveyor belt described by Carlson (1980), not all the trajectories rose rapidly as they moved to the west. In addition, we find no evidence that the airstream rises and turns anticyclonically towards the northeast. The trajectories presented in this study are more akin to the cyclonically turning, deep, closed circulation found in Schultz (2001) and are consistent with the cold conveyor belt trajectories described in Smart and Browning (2014) and Baker et al. (2014).
The trajectories that descended into the southwest wind-speed maximum from west of the cyclone are reminiscent of the cyclonic, dry airstream described by authors such as Danielsen (1964), Carlson (1980) and Young et al. (1987). These trajectories were associated with synoptic-scale descent within the cyclone (Figure 4(d)). They slowed between 1 and 21 m s$^{-1}$ as they descended, as a result either of friction, when they descended into the boundary layer, or movement against the local pressure gradient as the trajectories turned southwards (Figures 11 and 12(a–d, f)). However, trajectories in this airstream were still associated with winds of over 20 m s$^{-1}$ near the surface (Figure 12(a)).

Between the descending trajectories and the cold conveyor belt trajectories there was a set of transitional trajectories. These trajectories originated to the north of the low and initially rose as they moved westward around the low, but then descended into the southwest wind-speed maximum (Figures 11 and 12(b)). These transitional trajectories accelerated substantially on descent and resemble the sting jets described in Smart and Browning (2014) and Clark et al. (2005). The increase in wind speed of these transitional trajectories was a result of the along-flow pressure gradient force around the bent-back front.

Trajectories belonging to the cold conveyor belt experienced a larger along-flow acceleration from the pressure gradient force (up to 4 m s$^{-1}$h$^{-1}$) than the transitional trajectories because the cold conveyor belt trajectories were at lower levels where the pressure gradient was stronger (Figures 2(a, b) and 12(c)). However, because friction was also greater near to the surface, the cold conveyor belt trajectories (e.g. A12, B12) experienced weaker peak accelerations of 1–2 m s$^{-1}$h$^{-1}$ compared to 2–3 m s$^{-1}$h$^{-1}$ for the transitional trajectories that finished at a lower altitude (e.g. A20, B20) (Figure 12(d, f)). The trajectories that experienced
the greatest increase in wind speed did not experience the greatest positive along-flow pressure gradient force.

The idealised cyclone presented here exhibited two wind maxima, one to the east and one to the southwest of the low centre. Cyclones with two wind maxima were also found by Von Ahn et al. (2005) in a composite study of hurricane-force winds (33 m s$^{-1}$ or greater) in extratropical cyclones using QuikSCAT. Two wind maxima were also present in the Presidents’ Day cyclone of 1979 (Bosart, 1981). As in the idealised cyclone presented here, the southern wind maximum of the Presidents’ Day cyclone was co-located with a region of frontolysis (Bosart and Lin, 1984) and characterised by two airstreams: a descending, dry airstream and a cold conveyor belt (Whitaker et al., 1988). Indeed, Bosart and Lin (1984) noted the production of kinetic energy by flow down...
the pressure gradient north of the warm front and production of kinetic energy associated with descent south of the low. Moreover, Whitaker et al. (1988) demonstrated the acceleration of air parcels within the cold conveyor belt by the pressure gradient force north of the warm front. The similarity between the dry, idealised cyclone in this article and observations of real cyclones suggests that the life cycle of a dry, idealised cyclone may capture the essence of the wind field in real cyclones.

9. Conclusions

In this article, the strong winds within a dry, idealised baroclinic wave have been analysed using the horizontal momentum equation as the diagnostic framework. Two wind-speed maxima occurred in the simulation, one to the east of the low centre north of the warm front and the other to the southwest of the low.

Strong near-surface winds developed to the north of the cyclone underneath the warm-frontal zone. The flow was accelerated by the pressure gradient force at low levels, whilst continually under the effect of friction. This flow is similar to the cold conveyor belt described by Carlson (1980) and Schultz (2001), and developed because of the strong pressure gradient north of the warm front, which turned the background zonal flow and then accelerated the flow to the west with respect to the low centre. This airstream was responsible for the strong winds over 30 m s\(^{-1}\) to the east of the cyclone.

After passing through the eastern wind maximum, the flow either slowed or maintained a constant speed because of a decrease in the along-flow pressure gradient force around the north and northwest of the cyclone. The parcels moved around the bent-back front and were accelerated into the wind-speed maximum to the southwest of the low. Thus, the cold conveyor belt was also responsible for strong winds in excess of 35 m s\(^{-1}\) to the southwest of the cyclone.

Parcels from another airstream also entered the wind-speed maximum to the southwest of the low. This flow descended 100–300 hPa behind the cold front and brought high-momentum air from the mid-troposphere near to the surface. The trajectories in the descending airstream experienced a weak along-flow acceleration by the pressure gradient force until they moved southward and were slowed either by a negative along-flow pressure gradient force or by friction. Parcels from this airstream entered the east of the southwest wind maximum with winds of over 20 m s\(^{-1}\).

Between the descending airstream and cold conveyor belt, some transitional trajectories that resemble the sting jet described in Browning (2004) and Clark et al. (2005) also entered the southwest wind maximum. These trajectories accelerated whilst they descended, again as a result of a positive along-flow pressure gradient force, but above the boundary layer in the absence of friction. In particular, the occurrence of these transitional
traiectories in the absence of moisture suggests that airstreams resembling sting jets do not require the release of CSI or evaporative cooling.

We conclude that, despite the absence of moist processes, the simulation presented here captures the essence of the low-level wind field found in real extratropical cyclones.

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