Near-surface strong winds in a marine extratropical cyclone: acceleration of the winds and the importance of surface fluxes

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A rapidly developing extratropical cyclone named Tini brought strong winds to Ireland and the United Kingdom on 12 February 2014. A mesoscale-model simulation is used to analyze the development of the strong winds through the terms in the horizontal momentum equation. The maximum of near-surface wind speed equatorward of the cyclone was composed of two different airstreams that underwent different paths to acceleration. First, horizontally moving air in the cold conveyor belt was accelerated by the along-flow pressure gradient force but was decelerated by friction. Second, descending air accelerated into the eastern end of the maximum of near-surface wind speed and was associated, in part, with a sting jet, caused by the increasing along-flow horizontal pressure gradient force at lower levels. When this descending air entered the boundary layer, it too was decelerated by surface friction. Surface fluxes of heat and moisture were necessary to destabilize and deepen the boundary layer, allowing mixing of the strongest winds from the free troposphere down to the surface. A simulation with the surface fluxes turned off during cyclogenesis showed a more stable boundary layer around the bent-back front, which inhibited the strongest winds from reaching the surface. The descent of the sting-jet air was associated with a maximum in quasigeostrophic omega, which consisted of both synoptic-scale and mesoscale descent, the latter associated with frontolysis occurring at the end of the bent-back front. Thus, the near-surface wind maximum was created by the synoptic-scale and mesoscale dynamics, whereas localized moist processes were negligible.

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

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1. Introduction

To forecast strong winds in extratropical cyclones, Bergen School meteorologists would watch for ‘the poisonous tail of the back-bent occlusion’ (Grønås, 1995), a region of strong winds generally on the equatorward side of the low centre occurring at the end of the back-bent (or bent-back) front. As described in Schultz and Browning (2016), this concept was later adopted by Browning (2004), who described the wind pattern in the Great Storm of October 1987 and introduced the term sting jet into the meteorological lexicon. Browning found that the air comprising the sting jet started in the mid-troposphere poleward and west of the cyclone and descended cyclonically to the surface equatorward of the low centre. The publication of Browning (2004) motivated the growth in the UK of research on the causes and predictability of strong winds in the equatorward quadrant of extratropical cyclones. Although a few studies involving observational data (Parton et al., 2009; Martínez-Alvarado et al., 2014; Vaughan et al., 2015) and reanalyses (Hewson and Neu, 2015) have been performed since Browning (2004), most of the research within the last dozen years on strong winds in cyclones has been numerical modelling of real (e.g. Clark et al., 2005; Martínez-Alvarado et al., 2010; Schultz and Sienkiewicz, 2013) and idealized cyclones (Baker et al., 2014; Slater et al., 2015). Results from these studies include the following. Strong surface winds in cyclones occur within the warm sector, the cold conveyor belt poleward and west of the low centre and at the end of the bent-back front equatorward of the low centre (e.g. Hewson and Neu, 2015). Strong winds at the surface can also occur in conjunction with convective storms (e.g. Browning, 2004) or from tropopause folds (e.g. Parton et al., 2010). The wind maximum equatorward of the low is composed of at least two different airstreams: the cold conveyor belt and one or more descending airstreams corresponding to the sting jet (Martínez-Alvarado et al., 2014; Smart and Browning, 2014; Slater et al., 2015; Schultz and Browning, 2016). Air in the cold conveyor belt (Carlson, 1980; Browning, 1990; Schultz, 2001) travels cyclonically around the low centre at low levels, whereas air in the sting jet descends from the mid-troposphere (Clark et al., 2005; Baker et al., 2014; Slater et al., 2015).

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The physical process or processes responsible for the descending air and strong winds has been in question since Browning (2004). Browning (2004) hypothesized that the descent of the sting-jet air might have been caused by localized moist processes in the cyclone, specifically through one or both of two mechanisms: evaporative cooling and the release of conditional symmetric instability in moist slantwise convection (e.g. Bennetts and Hoskins, 1979; Emanuel, 1983; Schultz and Schumacher, 1999). The evidence for both of these processes in sting jets has been controversial. The process of localized evaporative cooling accelerating the flow has both supporters and detractors, with articles presenting evidence for (e.g. Clark et al., 2005; Browning et al., 2015) and against (e.g. Baker, 2014; Baker et al., 2014; Smart and Browning, 2014; Coronel et al., 2016) evaporative cooling being important to the acceleration of the winds. Similarly, the release of conditional symmetric instability accelerating the flow locally has also been addressed by articles presenting evidence for (e.g. Gray et al., 2011; Martinez-Alvarado et al., 2014) and against (e.g. Smart and Browning, 2014; Slater et al., 2015; Coronel et al., 2016) such a mechanism.

In contrast, a growing body of literature is indicating that synoptic and mesoscale dynamics are responsible for the strong winds. Schultz and Sienkiewicz (2013) and Slater et al. (2015) showed that the descent associated with the sting jet was related to an indirect circulation at the frontolitic end of the bent-back front. As the isentropes spread out during the weakening of the bent-back front, descent on the warm side of the front brings air down from aloft, which is then accelerated by the low-level along-flow pressure gradient. Coronel et al. (2016) found that the descent was primarily due to the along-front component of the Q vector, with a minor contribution from the across-front component (i.e. frontolysis). From a different perspective, Papritz and Schemm (2013) and Rivière et al. (2015a, 2015b) attributed wind maxima in the southern quadrant of the cyclones they studied to ageostrophic geopotential flux convergence (a term involved in the work provided by the pressure gradient force) along the bent-back front, also supporting the importance of synoptic and mesoscale dynamics for the acceleration of the wind. This body of literature and the controversy over the importance of localized moist processes leads to the conclusion that moisture is not an essential ingredient in the production of strong winds in a sting jet and that synoptic and mesoscale dynamics are sufficient. This conclusion – that local evaporative cooling is not a mechanism required to form the sting jet – is what we aim to test in this present article.

To calculate explicitly where the accelerations that led to the strongest winds were occurring in a cyclone, Slater et al. (2015) examined the terms in the horizontal momentum equation in a dry idealized model of a baroclinic wave. They found that the cyclone in such a simplified experiment was able to produce a sting-jet-like feature. The descending sting-jet air started out in a weak pressure gradient region in the mid-troposphere northwest of the cyclone. As it descended and moved cyclonically around the low, the air encountered the stronger pressure gradients in the lower troposphere to the west and south of the cyclone, which accelerated the air during its descent. As the air entered the boundary layer or started moving against the local horizontal pressure gradient, it stopped accelerating and often decelerated. Whether these results are also valid for strong winds in real extratropical cyclones that include moisture has not been tested. Therefore, the first goal of this article is to investigate the acceleration of strong winds in a real extratropical cyclone using the approach taken in Slater et al. (2015). In doing so, we aim to demonstrate that synoptic and mesoscale forcing is sufficient to explain the sting jet and that localized moist processes are negligible in this case.

Another result of Schultz and Sienkiewicz (2013) was that the sting jet occurred in a region of maximum sensible heat flux from the ocean surface. Schultz and Sienkiewicz (2013) suggested that the sensible heat and moisture fluxes could lead to convection, which deepens the boundary layer, which would in turn lead to higher momentum from above being mixed down to the surface. In contrast, without the fluxes, the boundary layer would be shallow and stable, inhibiting the mixing of momentum from the free atmosphere into the boundary layer. This hypothesis, however, has not been tested rigorously in a numerical model. Therefore, the second goal of this article is to investigate the modification of strong winds by surface fluxes in a model simulation of a rapidly developing cyclone.

The cyclone featured in this study brought strong winds to Ireland and the United Kingdom on 12 February 2014 (Figure 1). The storm left 115 000 homes in England and Wales without power and blew roof panels onto an overhead power line at Crewe railway station, resulting in the evacuation of 500 passengers (http://www.bbc.co.uk/news/uk-26153889, accessed 27 October 2014). This storm was chosen because it developed a region of strong winds southwest of the low in association with a comma-shaped cloud head (Figure 1). Furthermore, the storm developed over the ocean, like most cyclones that affect Europe, and will enable a better comparison with cyclones developing in idealized, baroclinic wave studies (e.g. Papritz and Schemm, 2013; Baker et al., 2014; Schemm and Wernli, 2014; Slater et al., 2015), due to the absence of large topography and land–ocean contrasts.

This study begins with an overview of the cyclone in section 2, followed by a description of the model set-up in section 3. A diagnostic framework to relate the generation of strong winds to local accelerations by individual terms in the horizontal momentum equation is introduced in section 4. The development of strong near-surface winds in the cyclone is then described in section 5. A Lagrangian perspective of the development of strong winds is given in section 6, where forces along air-parcel trajectories descending into the southwest wind maximum are calculated. In section 7, the effect of surface fluxes of heat and
moisture on the intensity of the low and the development of these near-surface strong winds is investigated. The origin of the descent is analyzed using the quasigeostrophic omega equation in section 8. Finally, our conclusions are presented in section 9.

2. Synoptic overview

Winter 2013–2014 was exceptionally stormy for the UK, being the wettest winter in England and Wales since 1766 (Eden, 2014). One of these storms was a low-pressure system named Tini by the Free University of Berlin, which crossed the UK on 12 February 2014. In advance of Tini, the Met Office released a Red Warning for hurricane-force winds in northern Wales and northern England, the first time a Red Warning had been released since January 2013 (http://www.metoffice.gov.uk/news/releases/archive/2014/Heavy-Rain, accessed 11 July 2016). Estimated insured losses from Tini were 286 million euros, mainly a result of damages in northwest England (http://www.perils.org/web/news/news-2015.html, accessed 19 February 2016).

From 10–13 February 2014, a strong zonal jet was present across the North Atlantic (Figure 2). Tini developed over the western Atlantic at 1800 UTC on 10 February, with an analyzed central pressure of 1004 hPa. This low-pressure system traveled eastward, south of a multi-low system centred at 58°N, 1004 hPa horizontal wind speed (m s⁻¹, shaded above 40 m s⁻¹) and 300 hPa geopotential height (red lines every 12 dam).

3. Model set-up

Tini was simulated using the Advanced Research Weather and Forecasting Model Version 3.4 (WRF-ARW: Skamarock et al., 2008). Global Forecast System (GFS) six-hourly, 0.5° analysis data from the National Oceanic and Atmospheric Administration (NOAA) National Centers for Environmental Prediction (NCEP) was used to initialize the model at 1200 UTC on 10 February 2014 and provide lateral boundary conditions. The model domain had a horizontal grid spacing of 20 km over a grid of 274 x 224 points ranging from about 30–65°N to 50°W to 10°E, with a model top at 50 hPa and 39 vertical levels, eight of those below 850 hPa. At the lower boundary, the Noah Land Surface Model was used with 10 min, 24 category US Geological Survey data (e.g. Ek et al., 2003; Tewari et al., 2004). The model was run with MM5 surface physics (Skamarock et al., 2008), the Thompson et al. (2008) microphysics scheme, the Kain–Fritsch convective parametrization scheme (Kain and Fritsch, 1990; Kain, 2004) and Yonsei University (YSU) planetary boundary layer scheme (Hong et al., 2006). The simulation also employed the Dudhia (1989) scheme for short-wave radiation and the Rapid Radiative Transfer Model (RRTM) scheme for long-wave radiation (Mlawer et al., 1997). This simulation is referred to as the control simulation (CNTL), which we later contrast with a simulation where the surface fluxes of heat and moisture are turned off (NOFLUX).

4. Momentum equation diagnostics

To analyze the development of strong winds in Tini, we employ the same horizontal momentum equation diagnostics as in Slater et al. (2015), but with two additional terms to account for explicit horizontal diffusion in the model (F^{DIFF} and curvature of the model grid (F^{CURV}). The horizontal momentum equation is formulated as

\[ \frac{\partial \mathbf{u}}{\partial t} = - (\mathbf{u} \cdot \nabla) \mathbf{u} - \frac{\partial p}{\partial x} \mathbf{v}_{\text{PK}} + \nabla \mathbf{HADV} - \nabla \mathbf{VADV} + \nabla \cdot \mathbf{F^{DIFF}} + \mathbf{F^{CURV}} + \mathbf{RES}, \tag{1} \]

where \( \mathbf{u} = (u, v) \) is the horizontal wind,

\[ \nabla \mathbf{h} = \left( \frac{\partial}{\partial x} + \frac{\partial}{\partial y} \right) \mathbf{h} \]

is the horizontal gradient operator, \( \partial / \partial t \) 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 eight terms on the right-hand side: \( \nabla \mathbf{HADV} \) is the horizontal advection of horizontal momentum, \( \nabla \mathbf{VADV} \) is the vertical advection of horizontal momentum, \( \mathbf{F^{COR}} \) is the Coriolis force, \( \mathbf{F^{PGF}} \) is the horizontal pressure gradient force, \( \mathbf{F^{BL}} \) is a term produced by the boundary-layer parametrization scheme, \( \mathbf{F^{DIFF}} \) is a term resulting from explicit horizontal diffusion in the model, \( \mathbf{F^{CURV}} \) is a term to account for curvature of the model grid and \( \mathbf{RES} \) is a residual term. The boundary-layer term \( \mathbf{F^{BL}} \) is calculated by taking the momentum tendency from the boundary-layer scheme and so includes the effects of friction and vertical mixing. The local acceleration \( \partial \mathbf{u} / \partial t \) is calculated from 15-min model output fields using a centred-difference method
Figure 3. Advanced Scatterometer (ASCAT) pass from about (a) 2200 UTC on 11 February and (b) 1145 UTC on 12 February. Wind is plotted (shaded according to the colour bar at the top right; a half-barb represents 5 knots, a full barb 10 knots and a pennant 50 knots) on a latitude–longitude grid labelled on the x- and y-axes. Images modified from the National Oceanic and Atmospheric Administration National Environmental Satellite, Data and Information Service. Available at http://manati.star.nesdis.noaa.gov/datasets/ASCATData.php.

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 residual term is mostly small across the domain (as will be seen in Figure 6 later). All other terms on the right-hand side of Eq. (1) ($HADV$, $VADV$, $F_{\text{COR}}$, $F_{\text{PGF}}$, $F_{\text{PBL}}$, $F_{\text{DIFF}}$, $F_{\text{CURV}}$) are output directly from the model. The contributions of the different terms in Eq. (1) to local changes in wind speed are calculated by projecting the forces in the horizontal momentum equation on to the unit vector of the horizontal wind.

A Lagrangian framework can also be used to analyze the forces acting on an air parcel. In this article, we focus on the relative contributions of the pressure gradient force and the boundary-layer term to the along-flow acceleration, because these two terms lead to the largest direct changes in wind speed. The Coriolis force acts perpendicular to the wind, so does not contribute to changing the wind speed. Horizontal advection, although quantitatively large, does not accelerate the wind in a Lagrangian framework. We also focus on vertical advection, because it facilitates the downward transfer of momentum toward the surface. For this purpose, air-parcel trajectories were calculated offline using the WRF software Read/Interpolate/Plot (rip; Staelinga, 2009). The trajectories were determined from model-derived velocity fields output every 15 min. The accuracy of the trajectories was improved by using a 3 min time step in rip (Staelinga, 2009).

The terms in Eq. (1) were interpolated along the trajectory from the 15 min model fields to the air-parcel location at each rip time step. Bilinear interpolation was used horizontally, whereas linear interpolation was used vertically and in time.

5. Acceleration of strong winds in the control simulation

The minimum central pressures of the cyclones from the control simulation (CNTL) and the Met Office surface analyses were in good agreement between 1800 UTC on 10 February and 0000 UTC on 13 February (Figure 4(a)). Two distinctive peaks in the domain-maximum total horizontal pressure gradient force occurred, one at 0600 UTC on 11 February and the other at 2345 UTC on 11 February, preceding maxima in the horizontal wind speed of 32 m s$^{-1}$ at 1100 UTC on 11 February in the warm sector and 42 m s$^{-1}$ at 0345 UTC on 12 February south of the low centre.

As in the analyzed cyclone, the simulated low in CNTL developed a bent-back front and exhibited a T-bone frontal structure (Shapiro and Keyser, 1990) by 0100 UTC on 12 February, when near-surface wind speeds greater than 35 m s$^{-1}$ developed in the strong pressure gradient southwest of the low (Figure 5(a)). The second peak in the domain-maximum total horizontal pressure gradient force occurred just after the most
advection $V_{ADV}$ also existed southwest and south of the low centre. The along-flow boundary-layer term $F_{PBL}$ generally followed the along-flow $F_{PGF}$ in magnitude but was opposite in sign.

Cross-section CD was constructed roughly along the major axis of the wind maximum at 0400 UTC on 12 February (Figure 5(b)). The cross-section reveals the vertical structure associated with the front, wind-speed maximum, accelerations, vertical velocity and frontogenesis (Figure 7). Between 0 and 250 km upstream of the wind maximum, the isentropes slope downwards to the southeast, indicative of the bent-back front. The magnitude of the along-flow horizontal pressure gradient force is largest near the surface and decreases upward upstream of the wind maximum, whereas the along-flow vertical advection term was maximum within the frontal zone (Figure 7(a)). A dipole in vertical velocity is evident, with a region of descending air to the west (about 0–100 km and 900–600 hPa in Figure 7(b)) and ascending air to the east (100–250 km and 900–600 hPa in Figure 7(b)). A prominent region of frontolysis lies at the eastern edge of the bent-back front, with descent on its warm side around 300–600 km above 800 hPa, consistent with the indirect circulation associated with frontolysis typically found at the end of bent-back fronts (e.g. Schultz and Sienkiewicz, 2013). This region of descent overlaps with some of the strongest winds in the mid-troposphere and also with the relative maximum in along-flow $V_{ADV}$. The isentropes here are near-horizontal, marking this as the region between the bent-back front and the cold front to the southeast. Other regions of large along-flow $V_{ADV}$ occur within the cross-section where ascending air transports momentum upward.

As will be further emphasized later, Figure 7 illustrates some key features of the cyclone structure and the accelerations...
and horizontal velocity (black arrows) are shown at crosses in Figure 8, represented by B10, D5, D10, E5 and E10 non-negligible only near the cold front.

Horizontal winds and potential temperature were smoothed with a four-point (red; solid is positive and dashed is negative) and boundary-layer term (brown, negative). The low-pressure centre is denoted by a light blue cross and ‘L’.

Components of the horizontal momentum equation for CNTL for Figure 6.

The air parcels started with speeds between 14 and 28 m s$^{-1}$ at 0400 UTC on 11 February, but then slowed to nearly zero wind speed by 2200 UTC (Figure 11). This deceleration was a result of a negative along-flow $F_{PGF}$ as the parcels moved to an area of weak horizontal pressure gradient to the north and northwest of the low (Figures 9 and 11(a,c,e)). After 2200 UTC on 11 February, the along-flow $F_{PGF}$ for these parcels became positive and increased quickly as they descended into a region of much stronger horizontal pressure gradient (Figures 11(b,c,e)). During the descent, the parcels accelerated to 42–48 m s$^{-1}$ in a mere 4 h (Figure 11(a)). This incredible rate of acceleration was tempered by the drag from friction between 0200 and 0400 UTC as the parcels descended closer to the ground (Figures 11(d) and (f)).

6. Lagrangian perspective on the control simulation

To gain a Lagrangian perspective on the strong winds in the control simulation, air parcels were released from every grid point in cross-section CD (Figure 8), with trajectories running backward for 24 h to 0400 UTC on 11 February. The minimum pressure that the parcels attained within these 24 h was subtracted from the initial pressure at 0400 UTC on 12 February to give a measure of the total descent of air parcels during this period. A value of zero means that the minimum pressure was attained at 0400 UTC on 12 February, implying that the parcel either ascended or achieved net zero vertical displacement in the preceding 24 h. Three regions of prominent descent (more than 200 hPa) are evident: beneath the bent-back front (0–150 km, 800–940 hPa), the main wind maximum (250–420 km) and within the cold front (550 km) (Figure 8).

Broadly, the trajectories occur in three groups and are similar to those in Slater et al. (2015). These groupings were obtained from a manual classification of all backward trajectories that ended in the cross-section in Figure 8. An easternmost group of trajectories ending within the cross-section (indicated as circles in Figure 8) resembled the warm conveyor belt. A second group (indicated as squares) accelerated at low levels as a result of the along-flow $F_{PGF}$ overcoming friction as they travelled around the western side of the low, similarly to the cold conveyor belt trajectories in Slater et al. (2015).

In this article, we concentrate on a third group of trajectories (crosses in Figure 8), represented by B10, D5, D10, E5 and E10

Figure 6. Components of the horizontal momentum equation for CNTL for 0400 UTC on 12 February. Horizontal wind speed (m s$^{-1}$, shaded above 20 m s$^{-1}$) and horizontal velocity (black arrows) are shown at $\sigma = 0.9965$ (approximately 25 m above the surface). Colours depict the 8 m s$^{-1}$ h$^{-1}$ contours of along-flow acceleration components for the pressure gradient force (blue), vertical advection (red; solid is positive and dashed is negative) and boundary-layer term (brown, negative). The low-pressure centre is denoted by a light blue cross and ‘L’.

Figure 7. Vertical cross-section CD through Tini from the CNTL simulation at 0400 UTC on 12 February. Cross-section location CD is shown in Figure 5(b). The vertical axis is pressure in hPa. (a) Potential temperature (thin black lines every 2 K), horizontal wind speed (m s$^{-1}$, shaded above 20 m s$^{-1}$), along-flow acceleration components contoured at 8 m s$^{-1}$ h$^{-1}$ (solid) and 16 m s$^{-1}$ h$^{-1}$ (dashed). The along-flow horizontal pressure gradient force is blue and the along-flow vertical advection term is red. (b) Petterssen frontogenesis (K (100 km)$^{-1}$ (3 h)$^{-1}$, coloured according to scale on right), potential temperature (thin black lines every 2 K), horizontal wind speed (m s$^{-1}$), black lines every 10 m s$^{-1}$) and vertical velocity (red = descent, blue = ascent, contoured at $-10$, $-5$, $5$ and $10$ cm s$^{-1}$). All fields were smoothed with a four-point horizontal window and a two-point vertical window.
These descending trajectories are similar to the transitional trajectories in Slater et al. (2015).

In both the present case and Slater et al. (2015), air parcels were accelerated by the along-flow horizontal pressure gradient force aloft in the absence of friction and then slowed slightly by friction as they entered the boundary layer. The parcels in this airstream have accelerations and descent rates comparable with those of other documented descending airstreams (e.g. Clark et al., 2005; Martinez-Alvarado et al., 2010; Baker et al., 2014; Smart and Browning, 2014), consistent with a strong, sting-jet-like airstream.

7. The effect of surface fluxes on the strong winds

To test the idea that the magnitude of the surface fluxes over the ocean affects the strength of the near-surface winds in Tinii, a simulation was performed with surface fluxes turned off after 1800 UTC on 11 February (NOFLUX). For comparison, surface sensible heat fluxes in the control simulation (CNTL) described in the previous section exceeded 400 W m$^{-2}$ southwest of the low centre near the time of strongest winds (Figure 12(a)).

As in CNTL, the NOFLUX cyclone continued to deepen rapidly after 1800 UTC on 11 February (Figure 4(b)). NOFLUX produced a similar cyclone but with a stronger cold front, less prominent bent-back front and a smaller area occupied by strong winds south of the low than in CNTL at 0400 UTC on 12 February (cf. Figures 5(b) and 13).

In the vertical, a prominent maximum in wind speed at 870 hPa is evident in NOFLUX, with wind speeds exceeding 50 m s$^{-1}$, contrasting with a maximum of 40–45 m s$^{-1}$ at this altitude in CNTL (cf. Figures 7(a) and 14(a)). A strong vertical gradient in wind speed in the lowest layers between 300 and 500 km along the cross-section was apparent, dramatically different to CNTL (cf. Figures 7(c) and 14(a)). These weaker near-surface winds in the absence of fluxes are consistent with reduced vertical mixing in the more stable boundary layer. This difference in static stabilities is illustrated by a comparison of the 950–975 hPa potential temperature difference (Figure 12). In CNTL, the atmosphere was statically unstable over much of the cyclone centre, including, crucially, the region of the wind maximum south of the low centre (Figure 12(a)). In contrast, instability in NOFLUX was limited to the region immediately behind the cold front (Figure 12(b)).

The patterns of along-flow vertical advection are markedly different in the two simulations (cf. Figures 6 and 15). In CNTL, the most prominent feature is the narrow arc-shaped region upstream of the wind maximum (Figure 6). Also present in Figure 6 is the separate region downstream of the wind maximum, parallel to and behind the cold front. In NOFLUX (Figure 15), along-flow $V_{ADV}$ is most prominent ahead of the wind maximum, with the main area of $V_{ADV}$ behind the cold front. This region is where the boundary layer is neutral or unstable in Figure 12(b).

The acceleration terms are largely similar between CNTL and NOFLUX (cf. Figures 7 and 14), except for along-flow $V_{ADV}$, where the main maximum in NOFLUX is lower than in CNTL and extends ahead of the wind maximum (Figure 14(a)). Because of the higher pressure gradient in NOFLUX north and west of the low centre, the region where the along-flow $F_{PGF}$ exceeds 8 m s$^{-1}$ h$^{-1}$ is broader and extends from northeast of the low centre (Figure 15). Countering this, however, is a much larger negative along-flow $F_{PBL}$ term in NOFLUX (Figures 6 and 15), a
result of a smaller positive contribution to along-flow $F_{PBL}$ from vertical mixing by the boundary-layer scheme in NOFLUX. Thus, turning off surface fluxes in the model simulation yielded the following results. Vertical advection and mixing was inhibited in the stable boundary layer upstream of the wind maximum in NOFLUX, leading to less acceleration overall than in CNTL and a reduced surface wind maximum (by 5 m s$^{-1}$). This result is consistent with the conclusions of Schultz and Sienkiewicz (2013) and Baker et al. (2014) that descent of strong winds to the surface southwest of the low centre is inhibited when the boundary layer is stable. In contrast, downstream of the wind maximum in NOFLUX where the boundary layer was unstable, along-flow vertical advection was enhanced. In the lower troposphere, along-flow vertical advection ahead of the lower-tropospheric wind maximum brought high momentum down. This downward transport of momentum offset friction to result in a smaller deceleration in wind speed downstream of the wind-speed maximum in NOFLUX than in CNTL. Frontolysis and downward motion along the bent-back front persisted longer in NOFLUX, leading to a wind maximum around 800–900 hPa not present in CNTL.

8. Synoptic-scale versus mesoscale forcing for descent in the lower troposphere

As further evidence that the descent of the sting jet is a result of large-scale dynamics of the cyclone and is not associated with localized diabatic processes, we calculate the forcing for descent around $T_{ini}$ using the quasigeostrophic omega. We use the Q-vector form of the quasigeostrophic omega equation formulated in pressure coordinates by Hoskins et al. (1978) and presented by Sanders and Hoskins (1990):

$$\sigma \nabla_p \frac{2}{f^2} \frac{\partial}{\partial p} \omega = -2 \nabla_p Q,$$  \hspace{1cm} (2)$$

where $\omega = \frac{dp}{dt}$ represents the vertical velocity in pressure coordinates, $\tilde{f}$ is the domain-averaged Coriolis parameter,

$$\sigma = -\frac{RT \frac{\partial \ln \theta}{\partial p}}{p}$$

is the effective static stability, $R$ is the gas constant of dry air,

$$\nabla_p = \left(1 \frac{\partial}{\partial x} + \frac{\partial}{\partial y}\right)$$

is the horizontal derivative on pressure surfaces and $Q$ is given by

$$Q = \left[\frac{\partial u_g}{\partial x} \cdot \nabla_p \left(\frac{\partial \Phi}{\partial p}\right), \frac{\partial u_g}{\partial y} \cdot \nabla_p \left(\frac{\partial \Phi}{\partial p}\right)\right].$$  \hspace{1cm} (3)$$

where $\Phi$ is the geopotential height. The quasigeostrophic omega $\omega_{QG}$ was calculated on pressure levels from Eq. (2) using

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Figure 11. Quantities calculated along selected 24 h backward trajectories taken from the wind-speed maximum at 0400 UTC on 12 February: (a) horizontal wind speed (m s$^{-1}$); (b) pressure (hPa); (c) along-flow horizontal pressure gradient force (m s$^{-1}$ h$^{-1}$); (d) along-flow friction force (m s$^{-1}$ h$^{-1}$); (e) magnitude of horizontal pressure gradient force and friction force (m s$^{-1}$ h$^{-1}$).
the method of successive over-relaxation, with a maximum error bound set at 0.0005 hPa s\(^{-1}\). The algorithm comes from Stoelinga (2009) and uses the domain-average Coriolis parameter and vertical stability profile. A lower boundary condition for calculating \(\omega_{QG}\) is determined from the model topography and an Ekman spiral.

The Q vector gives the vector rate of change of the potential temperature gradient following geostrophic flow, which indicates the direction of the secondary circulation that seeks to maintain thermal-wind balance. Regions of convergence of Q vectors in the lower troposphere are indicative of ascent, whereas regions of Q-vector divergence are indicative of descent.

To interpret the Q vectors further, we use the partition for vector generalization of the Pettersems (1936) frontogenesis function into vector components along the isentropes \(Q_s\) and across the isentropes \(Q_a\). This partition was introduced by Keyser et al. (1988), who used the total wind rather than the geostrophic wind in computing the vectors, whereas Barnes and Colman (1993) first applied the geostrophic-wind form for the case of a cut-off low over North America. The benefit of using this partition is that the divergence of the Q-vector component parallel to the isentropes is related to the component of the vertical motion on the scale of the baroclinic wave and the divergence of the Q-vector component normal to the isentropes is related to the component of the vertical motion on the frontal scale. Thus, this Q-vector partition can help test the idea proposed by Schultz and Sienkiewicz (2013) that the descent associated with the frontolysis (i.e. divergence of \(Q_s\)) at the end of the bent-back front was responsible for the descent of the sting-jet air.

Figure 12. Surface sensible heat flux (red lines contoured at 200 and 400 W m\(^{-2}\), potential temperature at 950 hPa minus potential temperature at 975 hPa (black lines at −0.5, −0.1, 0, 0.1, 0.5 and 1.0 K; solid for stable, dashed for unstable) and horizontal wind speed at \(\sigma = 0.9965\) (approximately 25 m above the surface; m s\(^{-1}\), shaded above 20 m s\(^{-1}\)), shown for (a) the CNTL simulation and (b) the NOFLUX simulation at 0400 UTC on 12 February. All fields were smoothed with a four-point horizontal window.

Figure 13. \(\omega_{QG}\) from the NOFLUX simulation at 0400 UTC on 12 February. Sea-level pressure (black lines every 4 hPa), potential temperature (red lines every 2 K) and horizontal wind speed (m s\(^{-1}\), shaded above 20 m s\(^{-1}\)) shown at the lowest model level \(\sigma = 0.9965\) (approximately 25 m above the surface). All fields are smoothed with a four-point horizontal window.

The 700 hPa Q vectors and \(\omega_{QG}\) at 0100 UTC on 12 February are compared with \(\omega\) from CNTL in Figure 16(a). The signs of maxima and minima are similar in both \(\omega_{QG}\) and \(\omega\), with ascent north and east of the low centre and descent southwest and south of it. Quantitatively, the largest difference occurs in the maximum of descent south of the low centre where the magnitude of \(\omega_{QG}\) of 8 Pa s\(^{-1}\) (−66 cm s\(^{-1}\)) is larger than the magnitude of \(\omega\) of 2 Pa s\(^{-1}\) (−16 cm s\(^{-1}\)), due to an overestimation of the horizontal wind speed by the geostrophic wind speed caused by the strong pressure gradient around the low at this location. This difference is also seen in previous studies of cyclones (e.g. Krishnamurti, 1968; Pauley and Nieman, 1992; Räsänen, 1996).

In the NOFLUX simulation, the frontolysis forcing from geostrophic motions produced a slightly larger \(\omega_{QG}\) descent southwest of the low centre than in CNTL (cf. Figures 16(a) and (b)). These patterns were similar regardless of the time of the simulation (not shown). These results indicate that the large-scale forcing for descent is similar in both the CNTL and NOFLUX simulations, which demonstrates that there is no large-scale difference between these two simulations and that the principal difference between them occurs in the boundary layer due to the surface fluxes only.

A closer look at the orientation of the Q vectors on the lowest model level shows that the Q vectors along the bent-back front were divergent and pointing to the northwest, nearly parallel to the isentropes (Figure 16(a)). In contrast, at the eastern end of the bent-back front, the Q vectors were divergent and pointing to the southwest, nearly perpendicular to the isentropes. The result is that there is a spatial difference between the descent associated with the baroclinic wave along the bent-back front and the descent associated with the frontolysis at the end of the front. This relationship holds for the NOFLUX simulation (Figure 16(b)) and also for other times in the evolution of the cyclone (not shown). This relationship is also present in the lower-tropospheric Q vectors in an idealized model of a marine extratropical cyclone with a sting jet in Coronel et al. (2016), their figure 13(a)). Both components of the divergence make up the strong descent region in both Coronel et al. (2016) and the present study, with the along-front component largest along the front and the frontolysis across-front component largest at the end of the bent-back front.

These results show that the quasigeostrophic descent in both the CNTL and NOFLUX simulations was more than adequate to explain the descent of the airstream responsible for the sting jet. Our results are similar to those of other studies that have found quasigeostrophic descent southwest of intense extratropical cyclones: Shutt (1990, his figure 6(c)) in the UK Great Storm.

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9. Conclusion

**Tini** was an intense marine extratropical cyclone that brought strong winds to Ireland and the United Kingdom on 12 February 2014. The results of our model simulations of **Tini** lead to the following conclusions.

1. The near-surface wind maximum south of the low centre was composed of more than one airstream. In particular, the air in the cold conveyor belt occurred to the west of air that descended from the mid-troposphere as part of the sting jet.

2. Multiple forces acted on the air as it accelerated into the wind maximum south of the low centre. In particular, this acceleration occurred as the parcels encountered a stronger along-flow horizontal pressure gradient as they descended around the bent-back front into the southwest wind maximum. As the descending air entered the boundary layer, friction began to offset the acceleration.

3. The descent was consistent with diverging Q vectors, indicating that the descent was caused by the dynamics.

Figure 14. Vertical cross-sections along **Tini** from the NOFLUX simulation at 0400 UTC on 12 February. Cross-section location **GH** is shown in Figure 13. The vertical axis is pressure in hPa. (a) Potential temperature (thin black lines every 2 K), horizontal wind speed (m s\(^{-1}\), shaded above 20 m s\(^{-1}\)), along-flow acceleration components contoured at 8 m s\(^{-1}\) h\(^{-1}\) (solid) and 16 m s\(^{-1}\) h\(^{-1}\) (dashed). The along-flow horizontal pressure gradient force is blue and the along-flow vertical advection term is red. (b) Petterssen frontogenesis (K (100 km\(^{-1}\)) (3 h\(^{-1}\)), colour according to the scale on the right), potential temperature (thin black lines every 2 K), horizontal wind speed (m s\(^{-1}\), black lines every 10 m s\(^{-1}\)) and vertical velocity (red for descent, blue for ascent, contoured at −10, −5, 5 and 10 cm s\(^{-1}\)). All fields were smoothed with a four-point horizontal window and a two-point vertical window.

Figure 15. Components of the horizontal momentum equation for NOFLUX for 0400 UTC on 12 February, horizontal wind speed (m s\(^{-1}\), shaded above 20 m s\(^{-1}\)) and horizontal velocity (black arrows) are shown at \(\sigma = 0.9965\) (approximately 25 m above the surface). Colours depict the 8 m s\(^{-1}\) h\(^{-1}\) contours of along-flow acceleration components for the pressure-gradient force (blue), vertical advection (red) and boundary-layer term (brown, negative). The low-pressure centre is denoted by a light blue cross and ‘L’. Horizontal winds and potential temperature were smoothed with a four-point window. Along-flow acceleration components were smoothed with a two-point window.

Figure 16. Horizontal wind speed (m s\(^{-1}\), shaded above 20 m s\(^{-1}\)) shown at the lowest model level \(\sigma = 0.9965\) (approximately 25 m above the surface) at 0100 UTC on 12 February, for (a) CNTL and (b) NOFLUX. Potential temperature (black lines every 2 K), omega (red, every 2 Pa s\(^{-1}\), solid positive, dashed negative), quasigeostrophic-omega (yellow, every 2 Pa s\(^{-1}\), solid positive, dashed negative) and Q vectors (black arrows sized to 1 m s\(^{-1}\) Pa\(^{-1}\)) at 700 hPa. All fields were passed through a nine-point smoother. Vertical velocity at this level in cm s\(^{-1}\) is approximately −8 cm s\(^{-1}\).
of the cyclone rather than local diabatic processes. Those diverging Q vectors revealed descent on the scale of the baroclinic wave along the bent-back front and descent caused by frontolysis at the end of the bent-back front.

(4) Surface fluxes of heat and moisture over the marine boundary layer facilitated mixing of the strong winds aloft into the boundary layer. When surface fluxes were turned off before the rapid development phase of the cyclone, the simulation was similar to the full-physics control simulation – in particular, the along-flow horizontal pressure gradient force accelerations around the cyclone. However, along-flow vertical advection into the southwest wind maximum was less in the no-flux simulation, due to the greater static stability in the boundary layer, resulting in surface winds that were about 5 m s\(^{-1}\) weaker than in the full-physics simulation.

Looking beyond this case study and linking these results to those of previous studies yields the following broader implications of this work for strong winds in extratropical cyclones, in general.

(1) The formation of sting jets are a response to the large-scale dynamics and do not require the local effects of evaporative cooling. Slater et al. (2015) and the present study both demonstrate that the strong winds are a response to the synoptic and mesoscale dynamics of extratropical cyclones. In the idealized cyclone in Slater et al. (2015), local moisture was not necessary to produce a sting-jet-like strong-wind feature. Schultz and Sienkiewicz (2013) and Slater et al. (2015) showed that the descent associated with the sting jet was related to an indirect circulation at the frontolytic end of the bent-back front. From a different perspective, Papritz and Schemm (2013) and Rivièrè et al. (2015a, 2015b) attributed wind maxima in the southern quadrant of cyclones they studied to ageostrophic geopotential flux convergence along the bent-back front, also supporting the importance of synoptic and mesoscale dynamics to the occurrence of the wind maximum. Indeed, the ageostrophic geopotential flux convergence is a term involved in the work provided by the pressure gradient force. Although some previously documented cyclones may have experienced greater near-surface wind speeds because of evaporating precipitation (Clark et al, 2005; Browning et al., 2015), this process is apparently not crucial to the formation of sting jets.

(2) Descending air in the sting jet potentially comprises the strongest wind in the cyclone, because the wind accelerates in the free troposphere where friction is weak. These results help explain why the momentum of air in the sting jet is so strong. Its acceleration occurs primarily above the surface in the free atmosphere during descent, unopposed by friction. In contrast, the air in the cold conveyor belt is always in the boundary layer and encounters friction, limiting its ultimate wind speed. This explanation is why sting jets are so potentially dangerous when they come on to land: the air descending into the boundary layer is highly accelerated air from aloft. Our results also echo those of Uccellini et al. (1987), who described the acceleration of the low-level jet in the February 1979 Presidents’ Day cyclone: ‘In effect, the rapid development of the [low-level jet] within 3 h represents a three-dimensional adjustment process in which parcels respond to changing pressure gradient forces due not only to horizontal displacement, but also to their vertical displacement within a baroclinic environment,’ (Emphasis is theirs).

(3) The stability of the boundary layer is crucial in determining whether the strong winds in the descending airstream get mixed down into the boundary layer and cause strong surface winds. Strong surface fluxes can deepen the boundary layer and entrain air from above. If this air were high-momentum descending air, strong and potentially damaging wind gusts could be brought to the surface.

With this last point, we see the opportunity for research on the conditions under which strong winds descend or are entrained into the boundary layer, both over the oceans and over land. Such research, combining both observations and modelling, would yield deeper insights into the problem of high winds in extratropical cyclones.

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