ROLE OF THE BOUNDARY LAYER

Topic 4.5

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Abstract: Progress on understanding the boundary layer is reviewed. Observations, idealised modelling and simulation with full NWP systems have all contributed to this progress. Attention has focussed on the characteristics of the turbulence, understanding the wind and thermal structure, how to model the boundary layer, how to parameterise the effects of the turbulence, and the flow dynamics. Substantial questions remain, including the nature and effects of the interaction between the boundary layer and the rest of the tropical cyclone.

4.5.1 Introduction

The boundary layer is an important part of tropical cyclones, since boundary layer processes regulate the sources of heat and moisture, and sink of momentum, which help determine the storm intensity, and because it is through the boundary layer that much of the impact on humanity and the environment occurs. Yet it has distinctly different properties from the atmospheric boundary layer elsewhere, because of the strong effect of the cyclone’s rotation upon its dynamics. While the importance of the boundary layer to tropical cyclone dynamics has long been recognised (e.g. Ooyama 1969), recent years have seen a surge of interest in the tropical cyclone boundary layer. This report will review progress on understanding the boundary layer since the last IWTC, although we mostly extend the review back to a little before 2010 in order to put the progress in better context.

To begin with, we must define the boundary layer. Textbooks typically contain statements such as (Kaimal and Finnegan 1994):

The boundary layer is the lowest 1 – 2 km of the atmosphere, the region most directly influenced by the exchange of momentum, heat and water vapour at the earth’s surface. Turbulent motions on time scales of an hour or less dominate the flow in this region, transporting atmospheric properties both horizontally and vertically throughout its depth.

This definition explicitly states that we should consider temperature, moisture and momentum in trying to determine where the boundary layer lies. Since the turbulence transports heat, momentum and moisture, there is usually good consistency between the boundary layer top derived from these three properties.
Observations of the tropical cyclone boundary layer, however, present a paradox. A composite analysis of a large number of dropsonde observations in the tropical cyclone core has shown that the inflow layer is about twice the depth of the constant-theta layer (Zhang et al. 2011b). That is, defining the boundary layer depth from wind data yields very different results to using temperature data. The same phenomenon was noted earlier from smaller numbers of observations by Wroe and Barnes (2003) and Schneider and Barnes (2005), as well as in numerical simulations using WRF by (Nolan et al. (2009a, b).

The difficulty of defining the boundary layer top in TCs has been discussed by Smith and Montgomery (2010), who have elsewhere defined the boundary layer as the layer of strong near-surface inflow which arises largely because of gradient wind imbalance (Smith et al. 2009). There is some evidence to suggest that the inflow layer is a better proxy than the well-mixed layer for the boundary layer in tropical cyclones. Measured vertical profiles of the moisture and momentum fluxes show that they approach zero not near the top of the well-mixed layer, as elsewhere in the atmosphere, but near the top of the inflow layer (Drennan et al. 2007; French et al. 2007; Zhang et al. 2008a). Momentum budgets from a boundary layer model of an axisymmetric cyclone similarly show that the momentum flux magnitude becomes small – specifically, it reduces to 20% of its near-surface value – near or slightly above the top of the inflow layer (Kepert 2010a, 2010c, 2013).

However, in a real (i.e. not axisymmetric) cyclone, the asymmetries may be strong enough that there is no inflow in some sectors, as in Hurricane Frederic (Powell 1980). Obviously there are still surface fluxes of momentum, heat and moisture in these regions, and hence a boundary layer. Thus identifying the boundary layer with the inflow layer has the difficulty that real storms are not axisymmetric (Kepert 2010a). Similarly, the supergradient flow in a moving cyclone can be quite asymmetric (Kepert 2001, Kepert and Wang 2001, Williams 2014) and so using the departure from gradient balance as an indicator of the boundary layer is also insufficient.

In this report, we will use a definition similar to Kaimal and Finnegan’s, except that we will allow the boundary layer to be shallower than 1 km. We note, however, that this definition cannot be applied without knowledge of the turbulent fluxes. If the data are limited to just wind, temperature and/or humidity, a proxy is required. At present, the question of how to detect the top of the boundary layer from such observations in a real (i.e. not axisymmetric) tropical cyclone does not seem to have a satisfactory answer.

4.5.2 Observations of the boundary layer (Jun Zhang)

The importance of turbulent processes in the boundary layer (BL) to tropical cyclone (TC) intensification and maintenance is well-known (e.g., Emanuel 1986; 1995; Smith et al. 2009; Bryan 2012; Zhang et al. 2012). However, the turbulence structure in the TC BL remains to be poorly understood, largely due to the lack of observational data. Till now, direct measurements of turbulence structure of the TC boundary layer over the ocean have only been taken by research aircraft with advanced turbulence sensors on board (Moss 1978; Black et al. 2007; Drennan et al. 2007; French et al. 2007; Zhang et al. 2008a; 2009; Zhang 2010). Most of these observations were taken from the Coupled Boundary Layers Air–Sea Transfer (CBLAST) field campaign in 2003–04 and are limited to the outer core region of a TC. Lessons learned from this the CBLAST experiment include:

1) Drag coefficients consistent with earlier results, showing a tendency to level off at surface wind speeds of ~25 ms\(^{-1}\);

2) Exchange coefficients for enthalpy fluxes are not dependent on the surface wind speed, up to speeds of 30 ms\(^{-1}\);
3) Turbulent fluxes of momentum, sensible heat and latent heat are nearly constant below 200m then decrease with increasing height in the hurricane boundary layer between the outer rainbands;

4) Turbulent fluxes tend to vanish near the top of the inflow layer and the height of the maximum wind speed not the top of the thermodynamic mixed layer;

5) Turbulent spectra of wind velocities and moisture are alike and tend to fall into the universal shape but temperature spectra have different shapes;

6) A turbulent kinetic energy budget showed unbalance between the shear production and dissipation term indicating horizontal advection may be important in TC BL dynamics.\(^1\)

A few eyewall penetrations within the boundary layer exist, such as during field experiments into Hurricanes Allen (1980) and Hugo (1989), but are rare for safety reasons. These data provided the quantitative estimate for momentum flux (Zhang et al. 2011a), in which the estimated maximum value of vertical eddy viscosity is \(\sim 100 \text{ m}^2\text{s}^{-1}\), much larger than used in the operational Hurricane Research and Forecast Model (HWRF). Increasing the vertical diffusion in HWRF lead to improved intensity, track and structure forecasts (Zhang et al. 2012).

Lorsolo et al. (2010) presented a novel technique to map turbulent kinetic energy (TKE) in several hurricanes using Doppler radar measurements. They found that TKE is larger mainly within the boundary layer and the whole troposphere of the eyewall region. This suggests that it is hard to define the top of the boundary layer using the TKE approach in the eyewall region (Smith and Montgomery 2010). Kepert (2012) noted that turbulence in the eyewall clouds is likely mainly generated in situ by buoyancy, rather than advected upwards from the boundary layer as suggested by Smith and Montgomery (2010). Given the limitation of turbulence observations in TCs, important scientific questions remain to be answered including:

1) What is the exact magnitude of enthalpy flux in the hurricane eyewall?
2) How do turbulent fluxes vary vertically in the hurricane eyewall?
3) Where is the top of the boundary layer using the flux approach?

The mean structure of the TCBL has been studied more than the turbulence structure, especially after the Global Positioning System (GPS) dropsonde data became available in 1997 (e.g., Franklin et al. 2003; Powell et al. 2003; Kepert 2006a,b; Barnes 2008; Bell and Montgomery 2008, Zhang et al. 2011b). The dynamics and kinematic structure of TCBL have been studied more than the thermodynamic structure using observational data. Franklin et al. (2003) are among the first who used dropsonde data to study the mean inner-core vertical wind structure of hurricanes. They found that the mean eyewall wind profile is characterized by a broad maximum centred at \(\sim 500 \text{ m}\), above which winds decrease with height. Below the height of the maximum wind speed, wind speeds decrease in the frictional boundary layer nearly linearly with the logarithm of the altitude, with the surface (10m) wind being approximately 75% of the peak value. Powell et al. (2003) analyzed the same data set as used by Franklin et al. and showed a logarithmic decrease in mean wind speed with decreasing height in the lowest 200 m, and studied the behaviour of drag coefficients. Kepert (2006a) used dropsonde data to study the boundary layer wind field of the core of Hurricanes Georges (1998) and Mitch (1998), and found that the flow was not supergradient in the former, but that the flow in the middle to upper boundary layer near the eyewall was strongly supergradient in the latter, with the imbalance being statistically significant (Kepert 2006a, b).

\(^1\) Although note that Kepert (2012), using a boundary-layer model with a higher-order turbulence closure, found that the advection of turbulence kinetic energy produced very little change in the simulations, including in the distribution of turbulence kinetic energy. Most of the turbulence was generated in situ by shear.
Model simulations confirmed this difference. Schwendike and Kepert (2008) extended the work by Kepert (2006a, b) to Hurricanes Danielle (1998) and Isabel (2003) and confirmed the strength of supergradient wind is related to the shape of the radial profiles of gradient wind and thus the inertial stability, again consistent with the simulations using the Kepert and Wang (2001) model. Bell and Montgomery (2008) also found supergradient flow in the boundary layer of Hurricane Isabel using dropsonde data. While the azimuthal-mean azimuthal flow in the upper boundary layer of most tropical cyclones is supergradient, the magnitude of the departure is quite variable, and in a few storms is smaller than is detectable or absent.

Zhang et al. (2011b) comaposited hundreds of GPS sondes from 13 hurricanes and studied the thermodynamic and kinematic structures of the hurricane boundary layer. Their results revealed that there is a clear separation of the BL height defined thermodynamically and dynamically, with the thermodynamic boundary layer height being much shallower than the dynamic boundary layer height. In particular, Figure 1 shows strong thermal stability above about 200-m height. This stable layer begins well within the inflow layer, which is about 800 m deep at the centre, increasing with radius to 2 km at 4 times the RMW. The height of the strongest azimuthal wind similarly increases with radius, and is substantially deeper than the layer of approximately constant $\theta$. That is, the thermodynamic boundary layer height and height of the maximum tangential wind occur inside the hurricane inflow layer. Stratification of the data shows that this holds regardless of hurricane intensity.

Figure 1. Radius-height sections of potential temperature (top), azimuthal wind (middle) and radial wind (bottom) composited from dropsonde data. Top panel, contour interval 0.5 K, 305 and 310 K thickened. Middle panel, contour interval 2 m s$^{-1}$, 40 and 50 m s$^{-1}$ thickened, dashed line shows the height of the maximum. Bottom panel, contour interval 2 m s$^{-1}$, 0 m s$^{-1}$ white dashed, black line shows 10% of the peak inflow.

From Zhang et al. (2011b)
Lee and Chen (2012) presented numerical simulations of asymmetric hurricane boundary layer structure in a fully coupled atmosphere-wave-ocean model and used these model simulations to compare aspects of the boundary layer structure against observations. They challenged the representativeness of the results given by Zhang et al. (2011b) and stated that the azimuthally averaged inflow layer tends to misrepresent the overall inflow layer structure TCs, especially the asymmetric structure. However, Zhang et al. (2014) argued that the methodology used by Lee and Chen (2012) is fundamentally flawed because the azimuthal mean structure which is only a function of radius and height is mathematically orthogonal to the asymmetric structure that is also a function of the azimuth. Zhang et al (2014) also argued that the simulated asymmetric boundary layer heights given by Lee and Chen (2012) essentially support the conceptual model of Zhang et al. (2011b). This conceptual model is also supported by the results of Zhang et al. (2013) who conducted analyses of the asymmetric boundary layer structure relative to the environmental shear using data from thousands of the dropsondes. They found that the separation of the dynamic and thermodynamic boundary layer heights and the decrease of boundary layer heights with decreasing distance are robust structures of the TCBL.

The asymmetric kinematic structure of the TCBL has been mainly studied using dropsonde data. Kepert (2006a,b) found that the ratio of the near-surface wind speed to that above the boundary layer increased inward toward the radius of maximum wind and was larger to the left of the track than to the right (in the Northern Hemisphere), and the low-level wind maximum is more marked on the left of the storm track than on the right (in the Northern Hemisphere). These observed structures are consistent with the asymmetric structure predicted by Kepert (2001) and Kepert and Wang (2001) using idealised diagnostic tropical cyclone boundary layer models. Zhang and Uhlhorn (2012) investigated the asymmetric structure of the surface inflow angle as a function of storm motion and intensity using thousands of dropsonde data. They found that the largest storm-relative inflow angles are found in the fastest-moving storms at large radii in the right-front storm quadrant, while the smallest inflow angles are found in the fastest-moving storms in the left-rear quadrant, for Northern Hemisphere storms. This observational result is also quantitatively consistent with the theoretical and numerical work given by Kepert (2001) and Kepert and Wang (2001). Recently, Barnes and Dolling (2013) analyzed data from 228 dropsondes in Hurricane Humberto (2001) and found that the asymmetric wind and mass fluxes are strongly tied to the storm motion and environmental vertical wind shear. They argue that convection bands near the storm centre produce a more complete wind field that offers greater protection to the warm core than the convective bands at larger radii, because the mass flux at the inner band correlates better with the storm intensity. This idea will hopefully be further investigated with additional case studies in the future.

As mentioned earlier, studies of the thermodynamic structure of the TCBL has been limited. Barnes (2008) investigated the thermodynamic structures of the TC lower-cloud and subcloud layers using dropsonde data collected in three hurricanes (Bonnie 1998, Mitch 1998, and Humberto 2001) and identified three unusual thermodynamic structures compared to those in a typical atmospheric boundary layer in a non-TC environment. First, positive lapse rates of equivalent potential temperature are found near the top of the inflow, which may insulate the inflow from the negative impacts of entrainment mixing from above. The second structure is a rapid decrease of equivalent potential temperature with height adjacent to the sea surface with a nearly superadiabatic lapse rate, which they suggested was because water loading from spray increases the residence time of air parcels in the surface layer. The third feature is a moist absolutely unstable layer found adjacent to the eyewall, in rainbands, and in the hub cloud within the eye. Recently, Zhang et al. (2013) studied the asymmetric boundary layer structure relative to the environmental wind shear using dropsonde composites. They found that \( \theta_e \) near the surface reached its minimum value left of shear (Northern Hemisphere), suggestive of evaporation-driven downdrafts bringing lower \( \theta_e \) from above. In the upshear right quadrant, the radial velocity is much
weaker in the boundary layer where boundary layer recovery starts via surface fluxes which eventually contribute to new convection initiating downshear right. They suggested that the boundary layer asymmetric thermodynamic structure and recovery processes are likely tied to shear-induced asymmetry of convection. This hypothesis remains to be tested in case studies where good coverage of thermal observations is available. How the thermodynamic structure is coupled with the kinematic structure and turbulent mixing in the TCBL will be also an interesting topic for future research.

4.5.3 Modelling the boundary layer

4.5.3.1 Winds (Jeff Kepert)

A range of models have been developed to improve understanding of TCBL winds, most of which aim to diagnose the boundary-layer flow in response to an applied pressure field representing that part of the cyclone above the boundary layer. They differ chiefly in the assumptions and simplifications made to obtain a solution. Kepert (2010a) categorizes these models based on the assumptions made:

- Column models: The one-dimensional single-column model of Deardorff (1972) was extended to the TCBL by Moss and Rosenthal (1975), further refined by Powell (1980) and Powell et al. (1996) and used for estimating surface winds from aircraft data.
- Depth-averaged (also known as slab) axisymmetric: Smith (2003), Smith and Vogl (2008). A simplified depth-averaged boundary layer model is an essential component of Emanuel’s potential intensity theory (Emanuel, 1988, 1995; Bister and Emanuel, 1998, 2002).
- Prescribed vertical structure, axisymmetric: Smith (1968), Leslie and Smith (1970), Bode and Smith (1975).

Some of these models have had considerable success in predicting and explaining observed features of the TCBL. For example, Franklin et al. (2003) analysed dropsonde data and found that the ratio of the surface wind speed to that at flight level increased from 0.8 at large radius to 0.9 near the radius of maximum winds (RMW), as predicted by Kepert (2001) and Kepert and Wang (2001). Similarly, those latter authors’ prediction of supergradient flow in the upper boundary layer in the tropical cyclone inner core was subsequently confirmed by observational analyses (Kepert, 2006b; Bell and Montgomery, 2008; Schwendike and Kepert, 2008). Further strong confirmation was provided by the model successfully reproducing the observational finding that the azimuthal-mean flow in Hurricane Georges was, unusually, not significantly supergradient (Kepert 2006a).

In contrast, other such models agree poorly with observations. For example, the axisymmetric depth-averaged (slab) model typically predicts that the frictional updraft near the RMW consists of either short-wavelength oscillations, or a complete breakdown in which the inflow becomes zero and the updraft becomes infinite. Figure 2 shows 2 sample simulations with this model which exhibit oscillations, a similar simulation with a shallower fixed boundary-layer depth produces the singularity (Smith and Vogl 2008). Which solution occurs depends on the parameters chosen, particularly the ratio $C_d/h$, where $C_d$ is the drag coefficient and $h$ the prescribed boundary-layer depth.
In an axisymmetric tropical cyclone, the fundamental effect of friction with the earth’s surface is to slow the near-surface azimuthal wind, disrupting gradient wind balance. Gradient wind balance is lost, and the agradient force (or gradient wind residual)

$$\text{AGF} = -\frac{v^2}{r} + f v - \frac{1}{\rho} \frac{\partial p}{\partial r}$$

is directed inwards, causing a cross-isobar flow component towards the centre of the cyclone. How strong is this inflow, and the consequent frictional convergence? Ooyama (1969) showed, under some simplifying assumptions, that the inflow is precisely strong enough so that the loss of absolute angular momentum $M_a$ to friction is replaced by radial advection (recall that in an inertially stable vortex, absolute angular momentum increases outwards, so that inflow implies that radial advection strengthens the azimuthal wind). In a cyclone with such a boundary layer, the boundary-layer flow cannot intensify the vortex because it imports only enough angular momentum to replace that lost to friction. Indeed, it should cause the vortex to decay because the outwards return flow, above the boundary layer, transports low-$M_a$ air from the centre of the vortex outwards.

Ooyama’s (1969) model depth-averaged the equations of motion, prescribed the boundary-layer depth, assumed that the azimuthal flow was in gradient balance and linearized the radial momentum equation. However, substantially less approximated models contain the same balance between frictional destruction and radial advection of $M_a$. In particular, the linearized model of
Kepert (2001) was shown by Kepert (2013) to produce this balance. That model preserves the height dependence of the equations, does not prescribe the boundary-layer depth, and does not prescribe gradient balance in the boundary layer; thus it imposes substantially weaker assumptions than Ooyama (1969). The chief assumptions made are the linearization (about gradient flow), and a simple parameterisation of the turbulent fluxes.

The first-order balance between frictional destruction and radial advection of $M_a$ implies that two further parameters are important. Firstly, increasing the surface drag, for example by increasing the drag coefficient $C_d$ or roughness length $z_0$, will clearly increase the rate of frictional destruction of $M_a$ and hence the inflow. Less obviously, a shallower boundary layer also leads to stronger inflow, because the loss of $M_a$ due to friction at the surface is balanced by radial advection integrated through the depth of the boundary layer. In the slab model equations, the drag coefficient appears mainly in the combination $C_d/h$, where $h$ is the boundary-layer depth, reflecting this fact. In the linearized height-dependent model of Kepert (2001), the depth scales as $\sqrt{2K/I}$ where $K$ is the vertical diffusivity and $I$ the inertial stability, so low $K$ gives a shallower boundary layer and stronger inflow. But in both of these first-order models, the frictional updraft at the top of the boundary layer is nearly independent of the boundary layer depth, because the updraft follows from integrating the continuity equation $\partial w/\partial z = 1/r \partial (ru)/\partial r$ through the depth of the boundary layer. In calculating $w$, the effects of the shallower boundary layer and stronger inflow compensate (Kepert 2013).

If we compare linear to nonlinear models, the terms omitted from the former in slab models are the radial advection, $u \partial u/\partial r$ and $u \partial v/\partial r$. Height resolving models add vertical advection, $w \partial u/\partial z$ and $w \partial v/\partial z$, and non-axisymmetric models add azimuthal advection, $v/r \partial v/\partial \lambda$ and $v/r \partial u/\partial \lambda$, to this list. We expect a shallower BL to be more nonlinear, because $u$ will be larger in magnitude, increasing the nonlinear radial advection. The vertical advection terms will also increase, even though to first order $w$ is similar, because the vertical gradient and hence the advection will be larger in a shallower boundary layer.

Smith and Vogl (2008) presented results with a nonlinear slab model, including examining the sensitivity of the flow to the imposed boundary layer depth. The results are somewhat consistent with the above discussion, in that the inflow is weaker when the boundary layer is deeper. The overall structure of $w$ changes little, but is displaced to larger radius when $h$ increases. However, the deeper boundary layer shows marked short-wavelength oscillations near the eyewall (such oscillations were first noted by Shapiro 1983 and first analysed by Smith 2003). On the other hand, with the shallower boundary layer the model became singular near the RMW, and shocks were formed. These shocks were subsequently analysed from the perspective of nonlinear waves and characteristics of the flow by Williams et al (2013), who attributed the formation of the very strong observed radial gradient in the boundary layer of Hurricane Hugo (1988) to this process (Marks et al 2008). However, we note that Marks et al. (2008), who attributed the gradient to an eyewall vorticity maximum, offer what appears a more convincing explanation, not least because the discontinuity and extreme updraft was absent on the opposite side of the storm.

Smith and Montgomery (2008) compared a range of approximations to the slab model, including linearization and Ooyama’s (1969) model (which they label Emanuel’s approximation, since it is also used in Emanuel 1986), to the otherwise unapproximated slab model. They found that most of the approximated models were similar to each other, but significantly different to the nonlinear model, and concluded that the terms omitted in the approximations were important.

One important issue with the use of slab models is the choice of drag coefficient. Drag coefficients are normally specified for the 10-m wind, but in slab models only the boundary-layer mean wind is available, which is of course stronger. So the drag coefficient should be reduced, as
otherwise the surface stress will be overestimated, leading to excessively strong inflow. Although this issue was noted by Ooyama (1969), it is sometimes ignored elsewhere (see the discussion in Kepert 2013). Increasing the drag coefficient increases the inflow strength, and hence the magnitude of the nonlinear terms. For example, Smith and Montgomery (2008) applied a 10-m drag coefficient to the boundary-layer mean wind for their calculations, likely affecting their conclusions about the relative magnitude of the nonlinear terms.

Continuing the theme of exploring nonlinearity in the tropical cyclone boundary layer, Vogl and Smith (2008) examined a partially linearized model which resolved the height dependence. They calculate the magnitude of the omitted terms, and conclude that they are not small relative to the retained terms, and cannot formally be neglected. Their results exaggerate the magnitude of the nonlinearity, because they used too small a turbulent diffusivity, $K = 10 \text{ m}^2\text{s}^{-1}$, leading to a very shallow boundary layer.

The difference between nonlinear and linear models of the TCBL had earlier been examined by Kepert (2001) and Kepert and Wang (2001), who showed that in the nonlinear model as compared to the linear model, there was an inwards displacement of the eyewall updraft, an increase in the extent to which the flow was supergradient, and an outflow layer above the peak supergradient flow. A further comparison, in the context of understanding the boundary-layer flow during eyewall replacement cycles, was presented by Kepert (2013). The differences were similar to those previously noted, but were found to be more marked near the outer RMW than the inner. Importantly, he showed that the two models were similar enough that the linearised model’s analytical solution was helpful in understanding important aspects of the nonlinear model simulations. (The role of the boundary layer in ERCs is discussed in section 4.5.6.2).

Montgomery et al. (2014), in commenting on Kepert (2013), noted a further limitation of linearized models such as Kepert (2001), in that they are highly sensitive to small-scale radial variations in the gradient wind. However, Kepert and Nolan (2014) noted that the diffusive time scale for the boundary layer is $1/I$ (Eliassen and Lystad 1977) and argued that this result implies that the nonlinear boundary layer only responds to gradient wind features above a certain length scale. They supported this scaling argument with calculations from both linearized and nonlinear height-resolving models.

In summary, nonlinearity is important in both depth-averaged (slab) and height-resolving (continuous) models of the TCBL. The effects seem to be larger in the slab model than in the continuous one, with slab models including several features that are contrary to observations, including shock formation, short wavelength oscillations, and frictional updrafts exceeding 10 m/s. Thus it is necessary to consider the differences between the height-resolving and slab models.

The differences between the height-resolving and slab models were considered by Kepert (2010a), who showed that the slab model produces much stronger inflow and a much greater departure of the boundary-layer mean winds from gradient balance. In moving storms, the slab models produce a much stronger motion-induced asymmetry, and can support an asymmetric analogue of the quasi-inertial oscillation previously noted in axisymmetric slab models. Kepert (2010b) used a new model, in which the vertical structure of the boundary layer is parameterised (rather than assumed constant or resolved) to show that the excessive inflow and agradient flow are due to excessive surface drag, while the tendency to oscillate is due to an inaccurate treatment

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2 This model differs from that of Kepert (2001) in that it does not linearise the surface boundary condition, thus an analytical solution is not available. It also omits the motion asymmetry.
of the vertically-averaged nonlinear terms. Kepert (2010b) also notes that the mechanism by which supergradient flow occurs is different between slab and height-resolved models.

When linearised, the differences between slab and height-resolved models diminish. Indeed, Kepert (2013) notes that the equation for the frictional updraft in the linearised height-resolving model of Kepert (2001) is similar to that in the balanced slab model of Ooyama (1969).

Ultimately, the question of which diagnostic boundary-layer model to use should depend on its ability to match the flow from more complete models (such as WRF) and from observations. The height-resolving linear and nonlinear models were compared to a WRF simulation of an eyewall replacement cycle by Kepert and Nolan (2014). The linear model is satisfactory, provided that the short-wavelength fluctuations in the gradient wind are first filtered out. The nonlinear model requires no such filtering, and agrees well with WRF, except that the boundary-layer inflow is about 10 – 20% too weak, presumably because of the omission of that component of the secondary circulation forced by latent heat release. The structure of the flow, including the location and relative magnitude of the eyewall updrafts, is well captured throughout the 48-hour period studied, encompassing the initial intensification, the secondary eyewall formation, intensification and contraction, and the replacement of the primary eyewall. In contrast, Abarca and Montgomery’s (2013) comparison of a similar full-model simulation to a nonlinear slab model calculation shows markedly poorer agreement.

The differences between slab and height-resolving models, both linear and nonlinear, full-physics models, and observations, are summarised in Table 1.

**Table 1. Typical values of boundary-layer flow parameters in various models and observations**

<table>
<thead>
<tr>
<th></th>
<th>Slab linear</th>
<th>Slab nonlinear</th>
<th>Height-resolved linear</th>
<th>Height-resolved nonlinear</th>
<th>Full-physics models</th>
<th>Observations</th>
</tr>
</thead>
<tbody>
<tr>
<td>Super-gradient flow</td>
<td>No</td>
<td>Yes</td>
<td>Slight</td>
<td>Yes</td>
<td>Yes</td>
<td>Yes</td>
</tr>
<tr>
<td>Oscillations</td>
<td>No</td>
<td>Often</td>
<td>No</td>
<td>No</td>
<td>Fine-scale spiral bands, but different process to nonlinear slab model</td>
<td>No evidence</td>
</tr>
<tr>
<td>Shocks</td>
<td>No</td>
<td>Often</td>
<td>No</td>
<td>No</td>
<td>Eyewall vorticity features, different process to nonlinear slab model</td>
<td>One example, but attributed to other causes</td>
</tr>
<tr>
<td>Inflow angle</td>
<td>Can exceed 45°</td>
<td>Typically 20 to 30°</td>
<td>Typically 20 to 30°</td>
<td>Typically 20 to 30°</td>
<td>Typically 20 to 30°</td>
<td></td>
</tr>
<tr>
<td>Typical azimuthal-mean eyewall updraft at top of BL</td>
<td>&lt; 1 m/s</td>
<td>10 m/s or more possible</td>
<td>&lt; 1 m/s</td>
<td>&lt; 1 m/s</td>
<td>&lt; 1 m/s</td>
<td>&lt; 1 m/s</td>
</tr>
</tbody>
</table>
4.5.3.2 Thermodynamics (Juliane Schwendike)

Plausible definitions for the tropical cyclone boundary layer include the inflow layer and the “well-mixed” layer, which is the layer over which the potential temperature is approximately constant. Observations (e.g. Zhang et al. 2011b) show that these two definitions give markedly different results, with the inflow layer being roughly twice the depth of the layer of nearly constant potential temperature. This discrepancy is both a hindrance to understanding, and has a potentially large impact on those boundary layer parameterizations that rely on the diagnosis of the boundary layer depth.

Kepert et al. (2014) derived the thermodynamics of the tropical cyclone boundary layer from the axisymmetric model CM1 (Bryan and Rotunno 2009). A marked thermal stability in the upper part of the inflow layer was found to be due partly to differential advection of potential temperature by the radial wind, which varies strongly with height, and partly to diabatic effects. The diabatic processes are associated with the condensation of moist air and the evaporation of rainfall into the subcloud air, and they lead to cooling near the surface, which diminishes with height. The differential advection of potential temperature by the radial wind contributes to the observed weak superadiabatic layer adjacent to the ocean surface, as the inflow has a maximum at about 100m height, and the potential temperature gradient is mostly directed inwards in the boundary layer. Thus cold air advection is maximized at about this height, with a weaker advective cooling tendency above and below. This term thereby stabilizes much of the boundary layer, except right near the surface and in regions where the radial temperature gradient is reversed. Since the stable layer is caused by processes within, rather than at the edge of, the boundary layer, it is not a suitable marker for the boundary layer top.

The effect of sea spray droplets, which are transported upwards by large eddies, on the thermodynamics and microphysics in the tropical cyclone boundary layer, was investigated by Shpund et al. (2014). They used a 2D hybrid Lagrangian model where the boundary layer was represented by the motion of an air volume towards a tropical cyclone eyewall along a background airflow.

Shpund et al. (2014) showed that subsidence and entrainment influxes into the tropical cyclone boundary layer regulate the budgets of humidity and temperature. In the absence of these influxes the humidity of the air rapidly increased and exceeded the saturation value, leading to the activation of the background aerosols, and to the formation of clouds containing small droplets and drizzle. Excluding the sea spray production did not produce a significant change in the thermodynamic structure of the boundary layer. Under the condition that the air humidity was relatively high, sea spray produced only drizzle drops with radii larger that 100 µm. When subsidence and entrainment influxes were not taken into account, specific humidity and potential temperature were larger than observed. In this case large droplets were mainly transported upwards. The role of sea spray markedly increased when the mean subsidence and the subgrid convective-scale influxes through the upper boundary layer were taken into account.

Shpund et al. (2014) found that the evaporation of sea spray lead to an increase in relative humidity of 10-15% and to a decrease in temperature by about 1-1.5K. Due to the high salinity of sea spray droplets, they grow under subsaturation conditions, which lead to the formation of drizzling clouds with a cloud base of about 200-250 m, and a radar reflectivity of about 5-10 dBZ.

4.5.3.3 Parameterisation (Akiyoshi Wada)

Simulated hurricane intensity and structure are affected by surface layer and planetary boundary layer (PBL) schemes that parameterize turbulent fluxes and vertical mixing processes in the boundary layer (Nolan 2009a,b; Kanada et al. 2012; Kepert 2010a,b; Smith and Thomsen 2010). A primary study reported differences of up to 15 m s$^{-1}$ in maximum wind speed and 16 hPa
in central pressure as well as those in simulated radar reflectivity patterns for Hurricane Bob (1991) among four PBL schemes including two local, one higher-order, and one nonlocal closure incorporated into the fifth-generation Pennsylvania State University-National Center for Atmospheric Research Mesoscale Model (MM5) (Braun and Tao, 2000). The comparison of the four PBL schemes using MM5 was studied by Smith and Thomsen (2010) in the framework of idealized numerical experiments.

Kepert (2012) further examined this problem. Unlike the earlier studies which used full-physics prognostic models and in which each simulation therefore had a different cyclone structure and intensity, Kepert (2012) used a diagnostic model in which each case had the same pressure field. Thus the differences found were due to the direct effects of the turbulence parameterisation, rather than to some unknown combination of the different parameterisation and different cyclone structure. Kepert summarized the formulation of the turbulent diffusivity in the four parameterizations used in tropical cyclone simulations, and categorizes the four parameterizations into:

- The Bulk and Hi-Res schemes
- Louis PBL scheme
- Nonlocal closures including the K-profile parameterization (KPP) schemes, medium-range forecast (MRS) and Yonsei University (YSU) schemes
- Higher-order closure including Mellor-Yamada (MY), Burk-Thompson (BT) and Gayno-Seaman (GS) schemes.

The Louis PBL schemes, nonlocal KPP closure and higher-order closures (MY, BT and GS) produced a logarithmic surface layer, while the Bulk and Hi-Res parameterizations were shown by both numerical experiments and mathematical reasoning, to fail to produce a logarithmic surface layer. The Bulk and Hi-Res parameterizations produced the strongest inflow, updraft and most strongly supergradient winds, and had the largest nonlinear terms in the budget equations, among the four categories.

Table 2 shows the number of studies that used at least one of PBL schemes belonging to the four categories and the Bulk and Hi-Res schemes are substantially the most popular schemes in MM5, representing 72% of the paper surveyed. The most popular PBL scheme is not always the "best scheme" for tropical cyclone studies. Kepert (2012) recommends the Louis PBL scheme and higher-order closures to produce the observed near-surface logarithmic layer (Powell et al., 2003). However, the recommendation was criticized by Smith and Montgomery (2013), who have questioned the validity of the traditional surface-based logarithmic layer in the inner core of tropical cyclones. They also noted that deviations from a logarithmic layer in the inner core of tropical cyclones must affect the ability to infer the surface drag coefficient from dropwindsonde wind profiles using methods that assume a logarithmic layer from the outset (Powell et al., 2003). We consider the logarithmic layer further in section 4.5.4.1.
Table 2. Boundary layer schemes used in articles that presented tropical cyclone simulations using MM5 or WRF in three prominent journals (Journal of the Atmospheric Science, Monthly Weather Review and Quarterly Journal of Royal Meteorological Society) published during 2006–10. Column heading abbreviations are Burk–Thompson (BT), Gayno–Seaman (GS), all Mellor–Yamada schemes (MY), medium-range forecast (MRF), Yonsei University (YSU), not specified (NS). The "not specified" category was ignored in calculating the frequencies. From Kepert (2012)

<table>
<thead>
<tr>
<th>Scheme</th>
<th>Bulk</th>
<th>Hi-Res</th>
<th>BT</th>
<th>GS</th>
<th>MY</th>
<th>MRF</th>
<th>YSU</th>
<th>NS</th>
<th>No. of papers</th>
</tr>
</thead>
<tbody>
<tr>
<td>MM5</td>
<td>10</td>
<td>32</td>
<td>3</td>
<td>1</td>
<td>7</td>
<td>10</td>
<td>—</td>
<td>6</td>
<td>64</td>
</tr>
<tr>
<td></td>
<td>17%</td>
<td>55%</td>
<td>5%</td>
<td>2%</td>
<td>12%</td>
<td>17%</td>
<td>—</td>
<td>—</td>
<td></td>
</tr>
<tr>
<td>WRF</td>
<td>—</td>
<td>—</td>
<td>—</td>
<td>—</td>
<td>—</td>
<td>13</td>
<td>2</td>
<td>37</td>
<td>7</td>
</tr>
<tr>
<td></td>
<td>—</td>
<td>—</td>
<td>—</td>
<td>—</td>
<td>—</td>
<td>27%</td>
<td>4%</td>
<td>77%</td>
<td></td>
</tr>
</tbody>
</table>

Experiments with a nonlocal KPP closure in a full-physics model produced the highest diffusivities, the weakest inflow, updraft and supergradient flow (Gopalakrishnan et al. 2013; Zhang et al. 2012). As higher diffusivity leads to a deeper boundary layer, the changes to the secondary circulation and supergradient flow in these studies are consistent with the effects discussed in section 4.5.3.1.

Recently, sensitivity of tropical cyclone models to the surface drag coefficients has been investigated in the framework of idealized numerical experiments (Montgomery et al. 2010) using four PBL schemes (Smith et al. 2014; Thomsen et al. 2014). Changing the drag coefficient provides insight into unbalanced effects in the boundary layer and their impact on the vortex evolution and structure. Thomsen et al. (2014) found that the intensification rate and mature intensity were essentially unaltered when the drag coefficient is perturbed randomly by variations of up to 60% so that they have questioned the notion that coupled wind-wave models are necessary to accurately forecast tropical-cyclone intensification and mature intensity. The result of numerical simulations for Typhoon Choi-wan (2009) by a nonhydrostatic model coupled with an ocean surface model and a multilayer ocean model showed that the difference in simulated central pressures was up to 10 hPa among five roughness-length schemes during the mature phase (Wada et al., 2013).

The solutions for a similarity model of the hurricane boundary layer are very sensitive to the specification of eddy diffusivity (Foster, 2009). Within the context of simple K theory, the average horizontal eddy diffusivity and mixing length are approximately 1500 m$^2$ s$^{-1}$ and 750 m, respectively, at about 500 m in the eyewall region corresponding to the mean wind speed of approximately 52 m s$^{-1}$ (Zhang and Montgomery 2012). The estimate of vertical eddy diffusivity determined from directly measured turbulent fluxes and vertical gradients of the mean wind speed, temperature and humidity supports using the KPP-type method to parameterize fluxes in the hurricane boundary layer for surface wind speed ranging from 18 to 30 m s$^{-1}$ (Zhang and Drennan, 2012). Improvements to the surface layer and boundary layer parameterization schemes in the operational HWRF based on in-situ aircraft observations (Figure 3) leads to the improvement of axisymmetric inflow layer structure (Figure 4) and reduction in hurricane forecast errors (Gopalakrishnan et al., 2013; Zhang et al., 2012).
1.5.3.4 Surface wind factor (Mark Powell)

For tropical cyclone basins with reconnaissance coverage, during the 1970-2000 era, forecasters needed to determine how best to adjust flight level measurements to estimate winds at the surface. Some of the early methods were empirically based, e.g. Powell and Black (1990), and also depicted the scatter of surface winds estimated from visual sea state observations. When the GPS sonde became available in the late 1990’s scientists examined the unique semi-Lagrangian wind profile measurements and were able to suggest relationships between the surface wind and the wind near typical flight levels (Franklin et al 2003, Powell et al., 2003). The Franklin et al 2003 paper found a suggested adjustment, or surface wind factor (R), of 90% of the flight level wind and this value was used to reexamine Hurricane Andrew of 1992 to reclassify the storm as Category 5 on the Saffir-Simpson scale.
The GPS sonde is limited for estimating peak surface winds due to the difficulty in launching the sonde at a location that will ensure it samples the surface wind maximum. During the past decade, a new instrument, the Stepped Frequency Microwave Radiometer (SFMR, Uhlhorn et al., 2007) has come on the scene which directly measures the emissivity of the ocean surface as increasing winds (and associated wave breaking) lead to white capping and foam layers within the water column. The SFMR overcomes the radial sampling limitation of the GPS sonde and performs well when corrected for rain (Klotz et al., 2014) and when used in strong enough winds to assure ample foam production.

Powell et al (2009) used pairs (surface, flight level) of measurements along the sloping radius of maximum winds from the SFMR and aircraft, respectively. Using Kepert’s and Kepert and Wang’s (2001) theoretical modeling of the boundary layer as a guideline, these data were examined to determine the variation of the ratio $R$ with radial distance, azimuth, storm motion, inertial stability, and angular momentum, resulting in regression equations for the maximum surface wind as a function of quantities determined from flight-level measurements. Their method explained much of the variance in $R$ because it accounted for the effects of storm structure on boundary layer winds, removed a bias of 4.6 m s$^{-1}$ and reduced the RMS error by 3.9 m s$^{-1}$ relative to the 90% rule. Franklin (2011) claimed that the bias helped make up for a perceived under sampling issue based on comparing a sampling pattern with a model simulation (later published as Nolan and Uhlhorn 2012). As noted by Powell et al (2011) and documented by Rotunno et al., (2009) and Bryan and Rotunno (2009), such wind fields are sensitive to available parameterizations, grid intervals, and tuning (mixing) parameters such that time step values are open to interpretation. Since SFMR is now available on both NOAA and Air Force reconnaissance aircraft, the new surface wind reduction factors are primarily a resource for revising the intensity of historical hurricanes that were sampled by aircraft before the advent of the SFMR.

4.5.4 Air-sea fluxes (Mark Powell)

As part of a string of important papers arising from the Coupled Boundary Layer Air-Sea Transfer (CBLAST) Experiment Zhang et al (2008a) used fast-response humidity, wind and temperature sensors to determine enthalpy fluxes at a series of altitudes in the boundary layer. While sensible heat flux showed an increase with decreasing height, humidity flux remains near constant with height. Based on these behaviours, an extrapolation to the surface was used to estimate the surface flux. Neither flux showed a wind speed dependence for the study range of 18-30 m s$^{-1}$. Based on measured ratios of enthalpy flux to momentum flux that are less than the theoretical value required to sustain a tropical cyclone (Emanuel 1995), the authors speculated on other potential sources of enthalpy flux, such as lateral fluxes through the eyewall (Persing and Montgomery 2003, Montgomery et al., 2006), or whether some alternative theoretical basis is applicable.

Sea state and drag coefficient behaviour continue to be topics of interest. Holthuijsen et al. (2009) analyzed a catalog of hurricane sea state images and found that active white cap coverage remained constant as winds increased, but total surface “whiteness” was increasingly dominated by streaks of foam, with sea spray also evident, until a “white out” was present in eyewall winds above 50 m s$^{-1}$. Following the profile method of Powell et al., 2003 to measure the surface roughness length, they found reduced roughness in extreme winds near the eyewall and higher values of roughness and drag in the outer left-front portion of the storm where cross swell is dominant. Once the winds increased beyond 40 m s$^{-1}$ in that quadrant, the roughness and drag decreased in similar fashion to other portions of the storm for the same wind speeds.
Bell et al. (2012) resurrected the budget study approach first used in a series of papers published by investigators with the National Hurricane Research Project during the 1960s, to estimate surface drag and enthalpy coefficients based on CBLAST data gathered in Hurricanes Isabel and Fabian of 2003. While the error bars involved in the budget technique are relatively large, they found no evidence of a wind speed behaviour for the enthalpy exchange coefficient and, consistent with the Zhang et al (2008a) study, they also found a ratio of momentum to enthalpy coefficients less than the theoretical value suggested by Emanuel (1995).

4.5.4.1 The logarithmic layer (Jeff Kepert and Mark Powell)

Smith and Montgomery (2014) have disputed the existence of a logarithmic wind layer near the eyewall in tropical cyclones. One objection they raise is that there is a significant change in wind direction, inconsistent with the nearly constant-direction flow expected in a logarithmic layer. However, the hodographs in their Fig 3 mark the wind direction change between the surface and 400-m height, which in some cases is deeper than the inflow layer. The logarithmic layer is only expected to be a good approximation in a relatively small part of the boundary layer – perhaps the lowest 10 to 20%. Expecting it to occupy most or all of the boundary layer is not a sensible comparison.

Smith and Montgomery (2014) present dropsonde observations from Typhoon Jangmi and Hurricane Isabel. Unfortunately, these data are binned to 50-m intervals and are plotted against linear, rather than logarithmic, height axes and so it is difficult to discern the details of the lower boundary layer flow. Figures 5 and 6 replots the same data against the logarithm of height, and it is clear that the wind speed profile is close to logarithmic from 20 to over 100 m height. Smith and Montgomery (2014) did present the observational composite data of Zhang et al (2011b) plotted against logarithmic axes, showing a marked near-surface departure of the wind speed from the logarithmic profile. We have examined a newer version of this dataset which includes additional observations and omits the filtering (as recommended by Nolan et al 2013). These data are plotted in Figure 7, and, in contrast to Smith and Montgomery’s (2014) similar figure, the logarithmic profile is clearly present.

Figure 7 shows a systematic departure from the logarithmic profile very near the surface. A similar departure is apparent in many previous depictions of such data (e.g. Powell et al. 2003). The cause of this is unclear, but we note that log-layer theory does not apply within several times the height of the surface roughness elements. Since a substantial part of the surface stress is carried by the waves, one would expect to see departures from the logarithmic layer up to perhaps 10 to 30 m from the surface in intense hurricanes.

Based on their analysis of the logarithmic layer, Smith and Montgomery (2014) also dispute Kepert’s (2012) analysis of boundary layer parameterisations for use in tropical cyclone modelling, and especially his recommendation that two popular parameterisations not be used because they do not produce a logarithmic layer. Interestingly, in Zhang et al (2014) these authors have criticised others for using one of these parameterisations, on the grounds that it does not produce a logarithmic layer. It is difficult to reconcile the views expressed in these two papers.
Figure 5. Eyewall mean wind speed from dropsonde data in Typhoon Jangmi on 24th (left), 25th (middle) and 27th (right) of September 2008 (red curves), together with the least-squares regression lines over data below 150-m height (blue line).

Figure 6. Eyewall mean wind speed from dropsonde data in Hurricane Isabel on 12 – 14 September 2003 (red curve), together with the least-squares regression line over data in the 20 – 150 m layer (blue line).

Figure 7. Plots of wind speed from the unfiltered dropsonde composite data similar to Zhang et al. (2011b) except not filtered, at 0.90 (right curve) and 1.08 (left curve) times the RMW (red curves), together with the least-squares regression lines (blue curves).
4.5.5 Small-scale wind structures (Sachie Kanada)

Small-scale wind structures, such as linearly organized coherent structures, are often observed in the tropical cyclone boundary layer (Morrison et al. 2005; Lorsolo et al. 2008; Zhang et al. 2008b; Ellis and Businger 2010; Koshioka and Wurman 2014). Recent works shed light on the influence of small-scale structures on surface-wind variation, vertical transport of momentum, and sensible and latent heat fluxes in the tropical cyclone boundary layer.

The coherent structures in the tropical cyclone boundary layer can be roughly divided into two categories: streaks dominate near-surface layers of the tropical cyclone boundary layer, including the surface and mixed layers (Lorsolo et al. 2008; Koshioka and Wurman 2014). Roll vortices are deeper features that span all or most of the boundary layer (Morrison et al. 2005; Zhang et al. 2008b; Ellis and Businger 2010). The temporal and spatial scales of streaks are shorter and smaller than those of roll vortices. The lifetime of streaks is measured in ten minutes, whereas that of roll vortices is measured in an hour.

Near-surface linear structures with subkilometer wavelengths (i.e., streaks) in the boundary layers of Hurricanes Isabel (2003) and Frances (2004) were investigated by using tower-based in situ measurement and land-based Doppler radar observation with a horizontal resolution of 50 m (Lorsolo et al. 2008). On average, the wavelengths in both storms ranged between 200 and 650 m. The features were in good agreement with the observation of Frances (2004) by two Doppler on Wheels mobile radars with a radial resolution of 25 m (Koshioka and Wurman 2014). Koshioka and Wurman (2014) reported that the wavelength was 400-500 m near the surface and the vertical flux of horizontal momentum was approximately 12 m² s⁻² in individual circulations which is much higher than the domain-wide average of the flux (2.5 m² s⁻²). Koshioka and Wurman (2014) speculated that this was likely due to the transient nature of the intense perturbations in the streaks. In general, the observation of streaks has been very few and limited on land, because it requires super-high temporal and spatial resolutions.

Zhang et al. (2008) reported the first in-situ aircraft-based measurement of roll vortices and roll-induced vertical fluxes in the tropical cyclone boundary layer of landfalling storms using the Coupled Boundary Layer Air-Sea Transfer (CBLAST) datasets. Zhang et al. (2008b) also employed the Canadian Space Agency’s RADARSAT satellite Synthetic Aperture Radar images and reported that mean wavelength of the rolls were approximately 900 m. In addition, Zhang et al. (2008b) investigated that the roll vortices in the tropical cyclone boundary layer increased the surface momentum fluxes by factor of 50% (Figure 8). Modulation of sensible heat flux by roll vortices at the middle level of the boundary layer was not significant, while that of humidity fluxes by roll vortices was relatively large (Figure 8). Morrison et al. (2005) reported previously that momentum fluxes (8 m² s⁻²) associated with roll vortices in the tropical cyclone boundary layer were about 2-3 times greater than those (2-3 m² s⁻²) estimated by standard turbulence models. The fluxes were larger than those reported in Zhang et al. (2008b). Zhang et al. (2008b) claimed the discrepancy in the momentum flux enhancement from Morrison et al. (2005) might be caused by their overestimate of roll wavelength (1450m) since Morrison et al. (2005) used WSR-88D radars with a radial resolution of 250 m.
Figure 8. Cumulative cospectral sums (ogives) of the along-wind (a) and cross-wind (b) components of the momentum flux (left) and the sensible heat (a) and humidity flux (b) (right) for NOAA43 boundary layer flight. The solid black line represents the cross-roll leg (A), the dashed black lines represent the three near-along-roll legs (B, C, and D), and the dash-dotted line represents the transverse-roll leg (E). The figures are adapted from Zhang et al. (2008b).

Another study using WSR-88D datasets in Guam reported roll vortices with a mean wavelength of 1350 m and momentum flux of 9.3 m$^2$ s$^{-2}$ associated with the passages of Typhoons Dale (1996) and Keith (1997) (Ellis and Businger 2010). Both values are almost the same as those in Morrison et al. (2005). The results also indicated an area decreasing roll coverage under the convective situations, suggesting that convective activity suppressed the genesis and/or maintenance of roll vortices.

Although, some discrepancies were found in the wavelength and enhancement rate of the flux, the observational studies showed the prevalence of coherent structures in the tropical cyclone boundary layer in both Atlantic hurricanes and Pacific typhoons during the landfall, and suggested the contribution of the coherent structures on the momentum fluxes in the tropical cyclone boundary layer. Meanwhile, French et al. (2007), who reported observations from Hurricanes Fabian (2003) and Isabel (2003) (major open ocean hurricanes of categories 4 or 5 on the Saffir-Simpson scale), indicated that there was no evidence of boundary-layer roll vortices in any of the data. The results indicate that there could be some typical condition either with or without roll vortices in the tropical cyclone boundary layer (Zhang et al. 2008b).

Large eddy simulations (LES) have been utilized to explore the structures, mechanism and roles of the coherent structures in the tropical cyclone boundary layer. Using the WRF–LES with a horizontal resolution of 100 m, Zhu (2008) investigated coherent structures in the tropical cyclone boundary layer and the associated vertical transport during the landfall of Hurricane Ivan (2004). The results indicated that the coherent structures in the tropical cyclone boundary layer existed in a mean stable environment and consist of well-defined updrafts and downdrafts. In addition, Zhu (2008) claimed that the current boundary layer schemes significantly underestimate the unresolved turbulent fluxes due to the fact that the effects of coherent structures in the tropical cyclone boundary layer had not been included in parameterizations (Fig. 9 in Zhu 2008). Zhu (2008) proposed a prototype updraft-downdraft model for parameterizing the fluxes induced by coherent structures in the tropical cyclone boundary layer.

Nakanishi and Niino (2013) conducted LES in the tropical cyclone boundary layer with a horizontal resolution of 40 m and studied roll vortices at two different radial distances from the storm centre, 40 km and 100 km, where the centrifugal force increased with decreasing radius. The EOF and budget analyses of turbulent kinetic energy demonstrated that an inflection point
instability in the radial velocity profile was responsible for roll vortices at both radii. Horizontal distributions of velocity fluctuations suggested the presence of streak-like structures at horizontal intervals of several hundred meters.

The dynamical height in the tropical cyclone boundary layer affects the wavelength of roll vortices (Gao and Ginis 2014). Following an approach of Ginis et al. (2004), Gao and Ginis (2014) showed that the wavelength of rolls increased as the dynamical height in the tropical cyclone boundary layer increased, while the growth rate of roll vortices decreased as the bulk shear decreased, under neutrally stratified tropical cyclone boundary layer (Fig. 9). They used a grid rolls-resolving model with horizontal and vertical resolutions of 32 m embedded in the basic-state tropical cyclone boundary layer model. The results also revealed that the mixing length of the model, indeed, had a great impact on the wavelength and growth rate of simulated roll vortices.

Figure 9. (a) Wavelength of rolls and (b) the dynamical HBL height scale (left), and (a) growth rate of rolls and (b) the bulk shear (middle) as functions of radius in Group N with different asymptotic mixing length (right).

The figures are adapted from Gao and Ginis (2014)

Rotunno et al. (2009) explored the effects of small-scale (<100m) three-dimensional turbulence in the vicinity of a tropical cyclone core by a multiple scale simulation. A horizontal resolution of the outermost domain was 15 km and that of the innermost nest was 62 m. As horizontal grid spacing reduced, large variations of winds suddenly appeared in the inner-edge of the eyewall, including filament-like fluctuating structures in 62-m grid. The maximum wind of 185-m to 62-m grid increased from 86 m s\(^{-1}\) to 122 m s\(^{-1}\), while the peak 1-minute mean wind and peak azimuthal-mean wind both decreased. These effects were caused by the larger turbulent diffusion in the 185-m grid than in the 62-m grid.

Rotunno et al. (2009) pointed out that dropsonde-derived winds of 107 m s\(^{-1}\) were observed in association with such filament-like fluctuating structures around the inner edge of the eyewall of intense hurricanes such as Isabel (2003). Meanwhile, by a virtual aircraft observation through a high-resolution numerical simulation of Isabel, Uhlhorn and Nolan (2012) found that the highest wind observed over a flight typically underestimated the 1-min averaged model wind speed by 8.5% ± 1.5%, though the 10-min averaged maximum wind speed was less underestimated (1.5% ± 1.7%) due to its corresponding larger spatial scale.

4.5.6 Gusts and Turbulence

Tropical cyclone winds are unsteady, with numerous gusts and lulls superimposed on the mean wind. It is often necessary to estimate the likely maximum gust given the mean wind, especially as damage relates more closely to gust than to mean wind speed. In addition, the unsteadiness of the wind implies that cyclone intensity will be different depending on whether it is measured as a 1-min, 2-min or 10-min mean wind. Harper et al. (2010) investigated the problem of
wind averaging period conversion, showed that the conversion factor should depend on the roughness of the underlying surface, and discovered that the previously used factors were appropriate for roughnesses representative of land rather than sea. They presented new recommendations, reproduced here as Table 3. The recommended at-sea factor for conversion from 1-min to 10-min averaging of 0.93 is substantially different from the previously accepted value of 0.88. These new factors have been adopted by several warning agencies, including France and Australia.

Harper et al. (2010) specifically highlighted the need to distinguish clearly between randomly sampled estimates of the mean wind speed based on any chosen averaging period and the peak gust wind speed of a given duration within a particular observation period. They particularly noted that mean wind speed measurements should not be converted between different averaging periods using gust factors, because a mean wind sampled at a fixed observing time is essentially a random sample. Its expected value is the true mean, and its actual value can fall either above or below the true mean, regardless of the averaging period. In contrast, a gust measurement is made by selecting the highest measurement from within a specified period, the selection making the sampling non-random. To emphasise, anemometer mean winds should not be converted for averaging period, nor should dropsonde measurements, and nor should satellite measurements such as passive radiometers or scatterometers.

Table 3. Cyclone intensity conversion factors for 1-min to 10-min winds, as recommended by Harper et al. (2010)

<table>
<thead>
<tr>
<th>$V_{max_{600}} = K V_{max_{60}}$</th>
<th>At-Sea</th>
<th>Off-Sea</th>
<th>Off-land</th>
<th>In-Land</th>
</tr>
</thead>
<tbody>
<tr>
<td>$K$</td>
<td>0.93</td>
<td>0.90</td>
<td>0.87</td>
<td>0.84</td>
</tr>
</tbody>
</table>

4.5.7 Impact of the boundary layer on the rest of the cyclone

4.5.7.1 Potential Intensity theory, sensitivity to $C_d$, $C_k$, and other turbulence parameters (Jeff Kepert)

Emanuel’s potential intensity (PI) theory (Emanuel 1986, 1988, 1995; Bister and Emanuel 1998, 2002) introduces the idea that the maximum intensity of cyclones depends on the ratio of enthalpy and drag coefficients, $C_k/C_d$. At the time, observations and theory suggested that at high winds, $C_d$ increased approximately linearly with wind speed, while $C_k$ was nearly constant, so that the ratio declined. This presented a conundrum, since the ratio became too small to support the observed intensity of tropical cyclones.

More recent observations (see section 4.5.4), showing that the drag coefficient is capped or possibly even declines above wind speeds of 25 – 30 m s$^{-1}$ and therefore suggesting that $C_k/C_d$ may be a little under 0.5 at high winds, somewhat alleviates this difficulty but does not completely resolve it.

The problem prompted a significant body of research on the effects of sea spray on the sea-air fluxes. Ultimately the question of spray-mediated fluxes is still unresolved, because the large observational deficiencies continue to frustrate the development of bulk parameterisations for these fluxes. A large part of the effect of sea spray is simply to repartition the enthalpy flux between heat and moisture, and whether the remaining component is critical for tropical cyclone development remains an open question.
Another attempt was to appeal to the boundary layer within the eye, where the low pressure allows very high values of $\theta_e$ to be achieved (Persing and Montgomery 2003).

Evidence against this hypothesis was provided by Bryan and Rotunno (2009), who showed that the effect of entrainment of this air into the eyewall only changes the intensity by $\sim 4\%$ on average.

The Emanuel PI model divides the cyclone into two parts: the boundary layer, and the rest of the cyclone. The latter part is taken to be in hydrostatic and gradient balance, with either reversible or pseudo-adiabatic moist thermodynamics. The boundary layer is modelled using a balanced slab model; in the context of section 4.5.3.1, this is a very simple boundary-layer model. While some recent criticism of Emanuel’s theory has centred on the boundary layer (e.g. Smith et al. 2008, 2009), such criticism has been heavily informed by results from nonlinear slab models, which are qualitatively less accurate than other commonly-used models (section 4.5.3.1 and table 1).

Rotunno and Bryan (2012) examined the sensitivity of PI to parameterised horizontal and vertical diffusion in an axisymmetric model. They found that the intensity, defined as the strongest azimuthal-mean wind in the storm (not necessarily at the surface) was mainly dependent on the horizontal mixing, with stronger mixing weakening the storm. However, the boundary layer structure and the extent to which the flow maximum was supergradient depended on the vertical mixing. Lower vertical diffusion led to a shallower, and more supergradient boundary layer, with stronger inflow and more "overshoot" (i.e. more nonlinear, consistent with the discussion in section 4.5.3.1). In such cases, the maximum gradient wind was lower, but a compensating increase in the degree to which the flow was supergradient nearly cancelled this out, leaving the peak velocity (i.e. the PI) with little sensitivity to the vertical diffusion.

4.5.7.2 The Role of Boundary Layer Dynamics in Secondary Eyewall Formation and Eyewall Replacement Cycles (Yi-Hsuan Huang, Jeff Kepert and Chun-Chieh Wu)

Unbalanced dynamics

The role of unbalanced flow within and just above the boundary layer top in secondary eyewall formation (SEF) was proposed by two recent companion studies (Wu et al. 2012; Huang et al. 2012; hereafter WH12). WH12 investigated the two mechanisms for the spin-up of azimuthal-mean tangential winds in single-eyewall tropical cyclones (TCs) highlighted in Smith et al. (2009). In WH12, two features were identified in the storm’s outer-core region around one day before SEF: 1) the horizontal broadening of low-level tropospheric swirling flow; and 2) intensification of boundary layer inflow. The association between increases in storm size and SEF was addressed from the axisymmetric viewpoint. The findings point to collective structure changes in the outer-core region of a mature TC, which ultimately culminates in the formation of a secondary eyewall. The sequence begins with the broadening of the low-level tangential wind field associated with the intensification of the eyewall that can be explained by the balanced response to radial gradient of diabatic heating (Schubert and Hack, 1982; Shapiro and Willoughby, 1982; Hack and Schubert, 1986; Smith et al. 2009). Due to the presence of surface friction, boundary layer inflow increases underneath the broadened swirling wind, and becomes large enough to enhance the swirling circulation within the boundary layer. The rapid increase in tangential winds leads to the local development of supergradient winds within and just above the boundary layer top, and thus to the decrease of boundary layer inflow. This process helps the organization of a boundary layer convergence zone where SEF subsequently occurs. Such a progressive increase in supergradient forces continuously provides a mechanical means for high-enthalpy air to erupt from the boundary layer. Given the dynamically and thermodynamically favourable environment for convective activities, the progressive responses of the unbalanced boundary layer flow to an expanding
swirling-wind field appear to be an important mechanism for concentrating and sustaining deep convection in a narrow supergradient-wind zone. While understanding the importance of the balanced response to diabatic heating, this study pointed out the important role of unbalanced dynamics in SEF.

A number of studies have examined this proposed idea from different perspectives. Concerning the asymmetry associated with rainbands that prevail prior to SEF, Qiu and Tan (2013) showed the impact of agradient forces on the downstream boundary layer flow, and subsequently on SEF. In addition, they indicated the important role of the pre-existing outer rainbands in SEF. Wang et al. (2013) investigated the depth-integrated boundary layer flow, and demonstrated that supergradient force and frictional force are the two leading terms contributing to the secondary maximum in boundary layer convergence. Sun et al. (2013) also found such an unbalanced flow characters associated with SEF, while emphasizing how the axisymmetrization of outer rainbands and the balanced response to peripheral heating in the rainbands lead to SEF (the latter mechanism was clearly demonstrated by Rozoff et al. 2012). By forcing a slab boundary layer with a flat tangential wind profile, Abarca and Montgomery (2013) argued that sole boundary layer dynamics was capable of enhancing secondary tangential wind maximum which is similar to results from a full-physics model simulation. Depicting secondary eyewall contraction, the result of the time-dependent slab boundary layer model also demonstrated the role of the unbalanced boundary layer flow in the eyewall replacement cycle. A limitation of their study is that a similar feature was previously identified as a artefact of slab models by Kepert (2010a). Deactivating all the model physics, except the planetary boundary layer process, Menelaou et al. (2014) carried out a WRF simulation in which the vortex continuously weakens and does not undergo SEF, and suggested that the boundary layer dynamics is not essential for SEF.

The unbalanced flow in the secondary eyewall has been revealed also by data collected during the Hurricane Rainband and Intensity Change Experiment (RAINEX; Houze et al. 2006) for Hurricane Rita (2005). Using dropsonde data winds at 500-m altitude, Didlake and Houze (2011) found apparent supergradient flow within both of Rita’s eyewalls. The collocation of the deceleration of the boundary layer inflow and supergradient forces was further demonstrated in Bell et al. (2012). Bell et al. (2012) also found that the alternating regions of convergence (the primary and secondary eyewalls) and divergence (the eye and moat) agree well with the radial distribution of the upward motion derived from the Electra Doppler Radar (ELDORA) data. It was also indicated that further quantitative analysis is needed to understand the relative importance of the unbalanced boundary layer processes in SEF.

Response of frictionally-induced upward motion to the radial gradient of vorticity

Taking a different perspective, Kepert (2013) used a nonlinear diagnostic boundary layer model to study the boundary-layer flow in a series of idealised vortices with perturbations in the wind profile at several times the RMW. Some of these perturbations were large enough to constitute discrete secondary wind maxima, but others were just small "bumps" in the profile. He found that these perturbations could cause relatively strong frictional updrafts; for example, if both gradient wind maxima were of similar strength, the updraft at the outer one was about twice as strong as that at the inner. Even small outer gradient wind bumps could create significant updrafts. Kepert (2013) explained the marked sensitivity of the friction-induced vertical motion to wind perturbations at outer radii using the updraft equation from a similar, but linearized, boundary-layer model. While there were some differences in the simulations between the two models (in the linear model, the updrafts were at larger radii, were more vertical, and were associated with weaker supergradient flow), the relative strength of the updrafts at the two RMWs were similar. The linear model provides a crucial physical link between the vorticity of the gradient wind and the frictional updraft. In particular, the updraft in the linear model is approximately proportional to the radial gradient of this vorticity, divided by the vorticity squared. Compared to the inner RMW, the outer
RMW has a weaker radial vorticity gradient, but is in an environment of much lower vorticity, so produces a relatively strong frictional updraft.

In his discussion, Kepert (2013) proposed a mechanism through which the boundary layer contributes to SEF and ERC through a positive feedback between vorticity perturbations, which produce an updraft through the above mechanism, that updraft favouring the development of convection, and the known tendency for convection to generate cyclonic vorticity (section 4.5.6.5). In that hypothesised feedback, the nonlinearity in the boundary layer was important, because it tended to place the frictional updraft at smaller radius, over the vorticity maximum, rather than outside of it.

![Figure 10. Hovmoller diagrams of the boundary-layer flow from a WRF simulation, together with the diagnosed flow using two diagnostic boundary-layer models. (a) 10-m radial flow from WRF, contour interval 2 m s$^{-1}$. (b) 10-m azimuthal flow from WRF, contour interval 5 m s$^{-1}$. (c) 1-km vertical velocity from WRF, contour interval 0.1 m s$^{-1}$. (d – f) as for (a – c), except according to the nonlinear BL model of Kepert and Wang (2001). (g – i) as for (a – c), except according to the linear BL model of Kepert (2001).]
Montgomery et al. (2014) proposed that Kepert’s (2013) hypothesis could be tested by comparing the actual updraft in a full-physics simulation of a SEF/ERC, with the diagnosed frictional updraft from a boundary-layer model. They found poor agreement, concluding they had falsified Kepert’s hypothesis. However, they used a linear model, in spite of Kepert’s clear statement that the nonlinearity was important to the hypothesised feedback. Kepert and Nolan (2014) presented similar calculations with a nonlinear boundary-layer model, and found excellent agreement, as seen in Fig. 10. They also showed that the strong sensitivity of the linear model to small perturbations in the gradient wind, which was the main reason for the poor agreement found by Montgomery et al. (2014), was absent in the nonlinear model. In short, Kepert’s (2013) hypothesis passes the test proposed by Montgomery et al. (2014), provided that the test is correctly performed.

Discussion
There are some significant similarities, and some significant differences, between these two points of view. Both agree that the developing outer eyewall has supergradient flow in the upper part of the inflow layer and a low-level inflow maximum. Momentum budgets from both are similar. Both agree that a wind field expansion is important, but Kepert (2013) notes in addition that it is important that it be nonuniform, in the sense of containing local maxima in the radial gradient of vorticity. But there are also some significant differences. Huang et al (2012) and Abarca and Montgomery (2014) attribute the deceleration of the frictional inflow and the development of the updraft to the outwards residual from gradient wind imbalance in the supergradient region. Limitations of their arguments are that they do not analyse all the terms in the radial momentum equation, and that the vertical distribution of the horizontal convergence does not coincide with that of the supergradient flow. In contrast, Kepert (2013) and Kepert and Nolan (2014) attribute the frictional updraft to the first-order balance of inwards advection of absolute angular momentum with its frictional destruction, together with nonlinear adjustment of the flow in the secondary eyewall region which displaces the updraft inwards (from the position given by the linear theory) and leads to the development of strong supergradient flow. That is, the supergradient flow is a byproduct, rather than a cause, of the frictional updraft. These latter authors argue that the boundary layer is slaved to the rest of the tropical cyclone, and that its flow can be well approximated as the response to the evolving pressure field at the top of the boundary layer. In contrast, Huang et al (2012) and Abarca and Montgomery (2014) argue that the time derivatives within the boundary layer are required. Kepert and Nolan (2014) further detail differences between the two viewpoints.

4.5.7.3 Impact of the Boundary Layer on the rest of the storm: Intensification (Chris Slocum)
Hurricane intensification has been described using axisymmetric, balanced theories like Wind-Induced Surface Heat Exchange (WISHE) (Yano and Emanuel 1991) and potential intensity (Emanuel 1988). These theoretical frameworks neglect the unbalanced dynamics typically observed in the hurricane boundary layer. Assuming that the boundary layer is in balance precludes supergradient winds from developing. More recent work has focused on the unbalanced dynamics, specifically relaxing the gradient wind balance approximation for the boundary layer to study the influence of low-level convergence of angular momentum. To understand the spin-up of the mean tangential winds due to boundary layer processes, Smith et al. (2009), Smith and Thomsen (2010), Montgomery et al. (2010), and Smith et al. (2014) conducted a series of experiments using a modified version of the Pennsylvania State University-National Center for Atmospheric Research fifth generation Mesoscale Model (MM5) to evaluate the boundary layer’s role in intensification.

Smith et al. (2009) identify two mechanisms for vortex intensification, both of which involve the radial convergence of absolute angular momentum. The first mechanism involves convergence above the boundary layer. The authors noted that modelled convergence resulted from radial buoyancy gradients developing in the presence of deep, inner-core convection and enhanced
surface moisture fluxes. This mechanism is consistent with previous work and can be interpreted in terms of balanced dynamics. The second mechanism is contained within the boundary layer of the tropical cyclone. The authors identified another region of convergence that is associated with the development of supergradient winds within the boundary layer. Smith et al. (2009) stated that the supergradient tangential winds act to decelerate the strong boundary layer radial inflow causing air to be carried upward and outward feeding the eyewall convection and resulting in storm intensification. The development of the supergradient winds cannot be captured by theoretical frameworks assuming gradient wind balance within the boundary layer. Accurately resolving the unbalanced dynamics of the boundary layer is key to understanding the tropical cyclone spin-up and intensification processes in their “new paradigm”.

To understand how the boundary layer is represented within 3-D numerical models, Smith and Thomsen (2010) evaluated the modelled tropical cyclone intensity by using five different planetary boundary layer parameterizations. The five schemes selected are the bulk, Blackadar, Burk-Thompson, MRF, and Gayno-Seaman. The authors found that the bulk and Blackadar schemes simulated a shallow inflow layer (800-850 m) with the largest inflow speeds and smallest radius of maximum wind. The MRF scheme simulated the deepest layer (1.5 km) with the weakest inflow and a large radius of maximum wind. The Burk-Thompson and Gayno-Seaman fell in between with inflow layers near 1 km. Smith and Thomsen (2010) found that drastic changes in boundary layer structure between the schemes are attributed to the eddy diffusivity values contained within each scheme. The authors stated that none of the schemes appear to capture a realistic boundary layer structure in terms of the radial inflow, supergradient winds, and vertical velocity. As a result, Smith and Thomsen (2010) did not recommend an optimum scheme for predicting tropical cyclone intensification.

Further investigation of boundary-layer parameterisations by Kepert (2012) revealed that two of the schemes, the bulk and Blackadar, were inconsistent with observations and theory because they did not produce a logarithmic surface layer. Based on this, and the fact that these schemes were originally formulated for the nocturnal boundary layer, Kepert (2012) recommended that they were not suitable for tropical cyclone modelling.

Montgomery et al. (2010) evaluated another aspect of how the boundary layer influences intensity of the tropical cyclone. In this case, the authors examined how intensity changes with the surface drag coefficient. Previous work shows that in an axisymmetric model, intensity decreases as the surface coefficient increases. Montgomery et al. (2010) found a different result. The authors’ simulations were conducted using a 3-D model framework. They found that intensity increases as the drag coefficient increases. The authors attribute the different findings to an increase in the agradient force within the boundary layer in the 3-D model. Montgomery et al. (2010) state that the increase in the agradient force results from the increase in frictional effects. Montgomery et al. (2010) compare their findings to potential intensity theory, which compares the ratio of enthalpy coefficient to the drag coefficient. The authors found that the ratio required for a tropical storm to intensify to a major hurricane is much smaller than the value proposed by Emanuel (1995). Bryan (2012) commented on the authors’ work by urging caution regarding the conclusions. Bryan (2012) states that Montgomery et al. (2010) did not run their model until a quasi-steady state was reached. If the simulation were run to the quasi-steady state, Bryan (2012) states that Montgomery et al. (2010) would have found a relationship between surface drag and intensity more consistent with results from axisymmetric model simulations. Bryan (2012) showed a result consistent with Emanuel’s (1995) enthalpy and drag coefficient relationships with simulations from CM1 run to a quasi-steady state. The conflicting results of Montgomery et al. (2010) and Bryan (2012) leave open the role of surface drag in intensification.
To further understand the role of surface drag in the intensification process, Smith et al. (2014) revisited the problem. This time, the authors chose four of the boundary layer parameterizations evaluated by Smith and Thomsen (2010): bulk, Blackadar, MRF, and Gayno-Seaman. The schemes are modified to use a consistent formulation of the drag coefficient. In the simulations, the values of moisture and enthalpy exchange are held constant. The value of the drag coefficient is halved and doubled from the accepted value to see how intensification of the vortex changes. The authors found a parameterization dependence on how storm intensity changes. However, the authors did see a similar trend across the schemes. Smith et al. (2014) show that intensity increased in all schemes as the drag coefficient is increased from the halved value to the accepted value then declines as the drag coefficient is increased from the accepted value to the doubled value (Figure 11). This result is a departure from the results found by Montgomery et al. (2010). Montgomery et al. (2010) found that intensity levels off as the drag coefficient continues to increase. As a result, the authors state that there exists an optimum drag coefficient associated with maximum intensity. Smith et al. (2014) also say that this result is still different than the relationship presented in the theory of potential intensity (Emanuel 1995). However, Smith et al. (2014) note that their work does not present a complete theory on how intensification is influenced by the unbalanced nature of the tropical cyclone boundary layer. They believe that more work must be done to understand the role the of the boundary layer in intensification and how to accurately model the effects.

![Figure 11. Radial profiles of time-mean azimuthally-averaged tangential wind speed at a height of 2 km for (a) the bulk and Blackadar schemes, and (b) the MRF and Gayno-Seaman schemes during the period 114-120 h. Curves 1, 2, 3 refer to runs with the halve the accepted drag coefficient, the accepted drag coefficient, and double the drag coefficient.](Figure 5 from Smith et al., 2014)

4.5.7.4 Impact of the Boundary Layer on the rest of the storm: Other storm properties (Chris Slocum)

Schubert et al. (1999) offered a simple dynamical explanation for mesovortices and other asymmetric features in the boundary layer by illustrating that the eyewall potential vorticity ring can be barotropically unstable. Mesovortices and other asymmetric dynamical features are often associated with localized regions of intense convection and wind. In the original argument made by Schubert et al. (1999), a 2-D barotropic non-divergent model was used. However, in current 3-D numerical simulations, turbulent processes that allow mesovortices to develop are not explicitly resolved and must be parameterized.

To understand how 3-D models simulate asymmetric structures and the formation of mesovortices, Zhu et al. (2014) varied the vertical turbulent mixing parameterization schemes using Hurricane Isabel (2003) as a case study. The authors’ objective was different from that of Kepert (2012), which focused on comparing numerical simulations using various vertical turbulent mixing parameterization schemes to observations.
Zhu et al. (2014) varied the Yonsei University scheme, the Mellor-Yamada-Janjic scheme, and the Mellor-Yamada-Nakanishi-Niino level-2.5 and level-3 turbulent kinetic energy schemes with initial and boundary conditions from the Geophysical Fluid Dynamics Laboratory (GFDL) model within the Advanced Research Weather Research and Forecast (WRF-ARW) model. The simulations of Hurricane Isabel (2003) were run for 48-h.

In their 3-D numerical simulations, Zhu et al. (2014) found significant differences in the simulations that influence eyewall asymmetric structures. The authors stated that the differences are the result of a complicated two-way feedback between the vertical turbulent mixing and diabatic heating resulting in generation or suppression of high- and low-frequency disturbances and in changes to the diabatic heating, intensity, and potential vorticity structure. As shown by the normalized power spectra of resolved vertical buoyancy, vertical total water, and horizontal radial PV fluxes (Figure 12), both the Yonsei University and the Mellor-Yamada-Janjic schemes generated high-frequency disturbances while both Mellor-Yamada-Nakanishi-Niino schemes were dominated by low-frequency disturbances. The findings are consistent with Nolan et al. (2009a,b) in that the high-frequency disturbances along the eyewall are produced with small eddy exchange coefficients in shallow turbulent layers and the low-frequency disturbances occur with large eddy exchange coefficients in deep turbulent layers, the latter of which eventually evolve into polygonal eyewalls.

![Normalized power spectra of resolved vertical buoyancy fluxes](image1.png)

Figure 12. Normalized power spectra of vertical buoyancy fluxes (a), vertical moisture fluxes (b), and radial potential vorticity fluxes (c) from 6 UTC 12 Sep 2003 to 0 UTC 14 Sep 2003 (Figure 12 from Zhu et al., 2014)
Zhu et al. (2014) did not attempt to quantify the realism of the asymmetric structures and mesovortices generated in the model simulations. With the development of current vertical turbulent mixing schemes having a mid-latitude origin, it still remains an open question as to which scheme can accurately represent these hurricane boundary layer processes to generate a more physically realistic simulation.

4.5.7.5 Impact of the Boundary Layer on the rest of the storm: Convection (Jeff Kepert)

Convection within tropical cyclones, as elsewhere in the atmosphere, may cause significant vorticity perturbations on the scale of the convection, which may persist for a significant period after the cloud which generated them has decayed. The more intense examples of this phenomenon, vortical hot towers (VHTs), were characterised by Hendricks et al. (2004) as being the preferred mode of convection, and by Montgomery et al. (2006) as generating a cyclonic vertical vorticity anomaly that is an order of magnitude greater than ambient. The primary focus of those initial papers was to explore the role of VHTs in cyclogenesis, although discussion of VHTs quickly developed elsewhere in the TC literature.

The main mechanisms by which convection generates vorticity are by stretching of ambient vorticity, by tilting of horizontal vorticity into the vertical, and by baroclinicity. The second of these processes implies that boundary layer structure is important, since the strong vertical wind shear in the boundary layer represents a substantial source of horizontal vorticity that convective updrafts can tilt into the vertical. Moreover, the complex structure of the boundary layer winds, including the low-level wind maximum, implies that the structure of the horizontal vorticity may be more complex than in midlatitudes.

A recent series of papers (Wissmeier and Smith 2011; Kilroy and Smith 2013; Kilroy and Smith 2013; Kilroy et al. 2014) have explored the development of vorticity within convection using a cloud resolving model and increasingly more complex background wind structures. Kilroy and Smith (2013) includes a low-level jet structure in which the horizontal vorticity vector reverses direction with height, and Kilroy et al (2014) includes a structure similar to that observed in tropical cyclones, in which the vector rotates with height. Both of these papers noted the complex vorticity structure in the resulting convection. While the convection had an overall tendency towards cyclonic vorticity due to stretching of the ambient vertical vorticity, it could also contain significant amounts of anticyclonic vorticity originating from tilting of horizontal vorticity from within the boundary layer, and the vorticity structure was quite different at different levels within the cloud. The vortical structure of the convection in these idealised studies is distinct from the VHTs as described by Hendricks et al (2004) and Montgomery et al (2006) because it includes vorticity of both signs.

Recently, Kepert (2013) and Kepert and Nolan (2014) have shown how the frictional updraft at the top of the boundary layer can be understood through the distribution of vertical vorticity within the boundary layer. Kepert (2013) proposed a positive feedback mechanism between frictional convergence, leading to stronger convection, leading to the generation of cyclonic vertical vorticity, leading to increased frictional convergence, and Kepert and Nolan (2014) presented an analysis of a simulated SEF and ERC which was consistent with this hypothesis (section 4.5.7.2). In this context, the recent studies of vorticity within tropical cyclone convection (Wissmeier and Smith 2011; Kilroy and Smith 2013; Kilroy and Smith 2013; Kilroy et al. 2014) are important because they confirm the link between convection and the tendency to preferentially generate cyclonic vorticity in a tropical cyclone environment.
4.5.7.6 Boundary layer changes at landfall (Jeff Kepert)

At landfall, a tropical cyclone experiences a marked change in the surface boundary, as land is generally rougher and drier than the sea surface, and following the evaporation of rainfall, cooler as well. During landfall, part of the cyclone circulation will be over sea, and part over land, for an extended period. Kepert (2006b) pointed out that the resulting frictional asymmetry is similar to that induced by storm motion, and showed that the observed wind asymmetry in Hurricane Mitch (1998) was consistent with this idea. Subsequently, two other case studies have explored this idea, May et al. (2008) in Severe Tropical Cyclone Ingrid, and Ramsay et al. (2009) in Severe Tropical Cyclone Larry.

Landfall causes changes to the gust characteristics of the flow, as well as to the mean winds. Zhu (2008) presented a simulation of the landfall of Hurricane Wilma during landfall, with the finest grid spacing of ~550 m capable of resolving convective downdrafts. He found that while the overall intensity decreased after landfall, there was no apparent trend in the gust strength because when the downdraughts of the circulations are in phase with the strong hurricane momentum aloft, they can effectively transport momentum downward to result in localized damaging winds. Larger land-surface roughness strengthened these boundary-layer secondary circulations by creating stronger local convergence, and thus led to stronger downdraughts and localized damaging winds despite the storm intensity weakening.

4.5.8 Summary

Substantial progress has been made in understanding the boundary layer in the past few years. Highlights are as follows.

- Further analysis of observations has confirmed that the upper part of the layer of strong near-surface inflow is quite stable, and shown that it is therefore difficult to define the top of the boundary layer. Observations and models show that the top of the inflow layer tends to correspond to the level at which the turbulent fluxes becomes small, and is hence a more appropriate definition, except for asymmetric storms. Analyses of model simulations have shown that the anomalous stable layer in the top of the inflow layer is due to diabatic processes and differential advection.
- Observational capability continues to improve. The SFMR instrument and CBLAST experiment have added to the already enormous contribution of dropsonde data in understanding of the boundary layer.
- Several further studies have confirmed Powell et al.’s (2003) finding that the drag coefficient is capped at high wind speeds. The enthalpy transfer coefficient has no detectable wind speed dependency. Progress has been made on understanding the implications of these results for potential intensity theory.
- Recent modelling of the boundary layer is divided between those who depth average the equations, and those who resolve the vertical structure. Linearized versions of these two classes are quite similar, but the full nonlinear versions have larger differences. Some of the behaviour of depth-averaged nonlinear boundary layer models is contrary to observations and “full-physics” models.
- Tropical cyclone simulations are sensitive to the parameterisation chosen. Which type of parameterisation is “best” remains unclear, but one popular class was shown to be unsuitable for use in tropical cyclones.
- Further evidence emerged of organised small-scale structures in the tropical cyclone boundary layer, including streaks and roll vortices. These contribute to surface wind gustiness and hence to damage, and to the turbulent fluxes within the boundary layer. The implications of this work for parameterisation are unresolved.
• A new study of gusts within tropical cyclones, incorporating a re-examination of the historical factors for adjusting cyclone intensity between 1-min and 10-min averaging, has led to new adjustment factors and a change in practice at some operational centres.
• Differing hypotheses on the role of the boundary layer in secondary eyewall formation have been proposed and analysed. One group of studies attributes the boundary-layer convergence to the existence of supergradient winds, but another shows that both the frictional updraft and the supergradient flow can be traced back to the structure of the gradient wind. In the latter view, the supergradient flow is not the cause of the updraft, but rather part of the process by which the boundary layer inflow adjusts to changes in the inertial stability.
• Boundary-layer structure can influence the distribution and sign of vorticity within tropical cyclone convection.
• The boundary layer structure changes at landfall for reasons including asymmetric surface roughness. These changes include asymmetric near-surface convergence.

References


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