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The influences of boundary layer mixing and cloud-radiative forcing on tropical cyclone size

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ABSTRACT

Tropical cyclone (TC) size is an important factor directly and indirectly influencing track, intensity, and related hazards, such as storm surge. Using a semi-idealized version of the operational Hurricane WRF (HWRF) model, we show that enabling cloud-radiative forcing (CRF) and enhancing planetary boundary layer (PBL) vertical mixing can both encourage wider storms by enhancing TC outer core convective activity. While CRF acts primarily above the PBL, eddy mixing moistens the boundary layer from below, both making peripheral convection more likely. Thus, these two processes can cooperate and compete, making their influences difficult to deconvolve and complicating the evaluation of model physics improvements, especially since the sensitivity to both decreases as the environment becomes less favorable. Further study shows not only the magnitude of the eddy mixing coefficient, but also the shape of it, can determine the TC size and structure.
1. Introduction

Tropical cyclone (TC) size is an important forecast metric as it directly and indirectly influences TC motion, intensity, track, and storm surge (e.g., Fiorino and Elsberry 1989; Fovell and Su 2007; Lin and Chavas 2012; Carrasco et al. 2014). There are a variety of metrics used to define the TC size, including the radius of the outermost closed isobar, the radius of vanishing wind, and the radius of 34-kt (about 17.5 m s\(^{-1}\)) wind speed (R\(_{34}\)). In this study, R\(_{34}\) at 10 m above mean sea level (MSL) is used to define the storm size or width.

Bu et al. (2014) demonstrated that cloud-radiative forcing (CRF), the interaction of hydrometeors with longwave and shortwave radiation, has an important role in expanding the storm size. Averaged through a diurnal cycle, CRF consists of pronounced cooling along the anvil top and weak warming through the cloudy air. In particular, the within-cloud warming was relevant, enhancing convective activity in the TC outer core, leading to a wider eye, a broader tangential wind field, and a stronger secondary circulation. This forcing also functions as a positive feedback (Fovell et al. 2016), assisting in the development of a thicker and more radially extensive anvil than would otherwise have formed. CRF itself depends on the microphysics parameterization and Fovell et al. (2010) showed it is a major reason why simulations can be sensitive to microphysical assumptions.

Bu et al. (2014) also demonstrated that the GFDL-derived radiation scheme that was long employed operationally in the Hurricane Weather Research and Forecasting (HWRF) model (cf. Tallapragada et al. 2014) did not handle CRF properly, resulting in deep clouds that were effectively transparent. Testing revealed, however, that implementing an ostensibly superior radiation scheme degraded model skill (L. Bernardet, M. Biswas, and C. Holt, personal communications, 2014). Analysis of those results led us to consider how the planetary boundary
layer (PBL) influences storm size, in cooperation and competition with CRF, which is the subject of this study.

It is widely appreciated that boundary layer processes play an important role in TCs (e.g., Smith 1968; Ooyama 1969; Emanuel 1986; Van Sang et al. 2008). Among these processes are mixing acting on momentum and scalars such as temperature and moisture, the subgrid portion of which can be represented via diffusion coefficients $K_m$ and $K_h$, respectively. Models such as HWRF typically handle vertical mixing within the PBL via a parameterization that presumes a local or non-local closure (cf. Stensrud 2007; Kepert 2012) such as the Mellor-Yamada-Janjić (MYJ; Janjić 1990) and Yonsei University (YSU; Hong et al. 2006; Hong 2010) schemes, respectively; Nolan et al. (2009b) provide a succinct comparison of these two approaches. Eddy mixing in local schemes depends on the specification of a vertical mixing length ($l_v$), stability, and either the vertical wind shear or prognosed turbulent kinetic energy. The HWRF model (see Tallapragada et al. 2014, 2015) has been using a non-local scheme called GFS (Global Forecast System) that, like YSU, is based on Troen and Mahrt (1986) and evolved from the Medium-Range Forecast (MRF) model’s PBL parameterization (Hong and Pan 1996).

There have been many papers addressing the sensitivity of simulated TCs to PBL parameterizations and assumptions (e.g., Braun and Tao 2000; Hill and Lackmann 2009; Nolan et al. 2009a,b; Smith and Thomsen 2010; Kepert 2012). Most previous studies of the PBL–TC relationship have focused on TC intensity, inner core convection, and/or the TC PBL structure. A few studies, however, have explicitly examined the influence of the vertical diffusion on the storm structure, most of them reporting no significant influence. For example, Bryan (2012) found the 34-kt wind radius to be only weakly sensitive (and the radius of maximum wind, or RMW, to be insensitive) to $l_v$, at least when reasonable values of the horizontal mixing length ($l_h$) and the enthalpy/drag coefficient ratio ($C_k/C_d = 0.5$) are used. Bryan and Rotunno (2009) and Chavas and
Emanuel (2014) both demonstrated that the RMW remained essentially unchanged with doubled $l_v$. While Frisius (2015) did find some sensitivity to vertical diffusion, he pointed out that the lifetime of real TCs is too short for the effect of vertical diffusion to become relevant.

With the HWRF model, we demonstrate herein that vertical mixing can exert a very important influence on storm size within several days after initialization, especially when the operational GFS PBL scheme is employed. In that scheme, the PBL depth, $h$, is determined using an iterative bulk-Richardson approach calculated from the surface upward. The profile of vertical eddy diffusivity applied to momentum between the surface and $h$ is obtained via

$$K_m = \kappa \left( \frac{U_*}{\phi_m} \right) Z \left[ \alpha \left( 1 - \frac{Z}{h} \right)^2 \right],$$  

where $\kappa$ is the von Kármán constant ($=0.4$), $U_*$ is the surface friction velocity scale, $\phi_m$ is the wind profile function (nondimensional shear) evaluated at the top of the surface layer, and $Z$ is the height above the surface. This formula produces a mixing coefficient profile that is parabolic in shape between the surface and $h$ with a maximum at $Z = h/3$. The vertical eddy diffusivity for temperature and moisture, $K_h$, is obtained by dividing $K_m$ by the turbulent Prandtl number, $Pr$ (Hong and Pan 1996). For the GFS scheme as implemented in HWRF, $Pr \approx 1$ within the hurricane PBL, making $K_h$ approximately equal to $K_m$, but the handling of this proportionality differs even among closely related schemes such as YSU (Hong and Pan 1996; Hong et al. 2006; Hong 2010).

Gopalakrishnan et al. (2013) demonstrated that eddy mixing strongly influences the intensity and depth of the TC low-level inflow and the GFS PBL parameterization was producing excessively large $K_m$ values relative to those estimated from observations by Zhang et al. (2011). They addressed this shortcoming of the GFS scheme by incorporating a tuning parameter, $\alpha$, into

\[ K_m = \kappa \left( \frac{U_*}{\phi_m} \right) Z \left[ \alpha \left( 1 - \frac{Z}{h} \right)^2 \right], \]

1Additionally, according to the WRF code version 3.7.1, YSU applies a separate mixing coefficient, $K_q$, to moisture, which in the mixed layer utilizes a somewhat modified turbulent Prandtl number formulation from that employed for $K_0$. This results in slightly different mixing being applied to moisture and heat.
While a value of $\alpha = 0.25$ was found to produce the most reasonable results relative to the observations at 500 m MSL for wind speeds in the range 10–60 m s$^{-1}$, a setting of $\alpha = 0.7$ was adopted in the 2013 and 2014 operational versions of HWRF (Bernardet et al. 2013; Holt et al. 2014) in all three of its telescoping domains as a consequence of skill testing against retrospective TC cases. However, this may have introduced a mismatch between the PBL and surface schemes when $\alpha \neq 1$ that, while addressed in the 2016 HWRF operational model (Weiguo Wang, personal communication, 2016), persists in the version employed herein. The influence of $\alpha$ on the near-surface wind and vertical shear and its consequences are explored in Sec. 3b(3).

In this study, we focus on uncovering how and why the PBL vertical mixing impacts horizontal TC structure and size. The model and experimental design are discussed in Section 2. Section 3 demonstrates how and why CRF and PBL mixing cooperate, and compete, to influence TC size. Section 4 presents the summary discussion.

2. Model and experimental design

The HWRF simulations in this study were carried out using the 2014 operational code. These experiments are “semi-idealized” in that we simplified the operational configuration by excluding land and decoupling the ocean model, employing a uniform and constant sea-surface temperature (SST) of 302.5 K for the standard runs, and initializing with a horizontally homogeneous tropical sounding [modified from Jordan 1958; see Fovell et al. 2010] without any mean flow. The Cao et al. (2011) “bubble” procedure was used to initiate the TC and all simulations spanned four full days with composite model fields being constructed for the final day in a vortex-following fashion, averaging over one full diurnal cycle. While changes in operational settings (primarily with respect to horizontal smoothing) from the 2013 version used by Bu et al. (2014) resulted in somewhat weaker storms for the same experimental design, all of the standard SST HWRF TCs attained
major hurricane status (at least Category 3 on the Saffir-Simpson scale) for the analysis period.

Please note that most figures employ azimuthal averaging, thereby understating the maximum
intensity of these asymmetric, beta-sheared storms (cf. Bender 1997; Bu et al. 2014).

As in Bu et al. (2014), our simulations employed three telescoping domains (with 27, 9, and 3 km
horizontal grid spacings) along with some of the model physics used operationally during the 2014
season, such as the SAS (Simplified Arakawa-Schubert) cumulus parameterization (remaining
active in the 27 and 9 km domains after 24 h). In 2014, the operational configuration (cf.
Tallapragada et al. 2014) also included the GFDL radiation scheme, the GFS PBL, and the tropical
Ferrier microphysics parameterization (MP). These are compared to (or replaced by) RRTMG
radiation (Iacono et al. 2008), the YSU PBL (Hong 2010), and Thompson MP (Thompson et al.
2008), respectively. Bu et al. (2014) showed that while RRTMG and GFDL generate nearly
identical clear-sky radiative forcing profiles owing to longwave and shortwave radiation (see their
Fig. 7a), CRF was not properly handled in the HWRF implementation of the latter. Therefore,
owing to their strong similarity with RRTMG cases in which cloud-radiative forcing is deactivated
(cf. Bu et al. 2014), HWRF simulations with GFDL radiation are also labeled “CRF-off” herein.
Our work has shown that storm structure is significantly modulated by microphysical assumptions
(cf. Fovell et al. 2016), but for simplicity we will focus solely on the Thompson scheme.

Also following Bu et al. (2014) we employ an axisymmetric version of Cloud Model 1 (CM1)
model (Bryan and Rotunno 2009) initialized with the Rotunno and Emanuel (1987) sounding.
These simulations used 3 km radial grid spacing and 100 m resolution in the vertical (below 4
km MSL), were initialized with a weak vortex, and were integrated for twelve full days. As in Bu
et al. (2014), Goddard radiation (Chou and Suarez 1994) and a version of Thompson microphysics
were used for most experiments and the latitude was 20°N. Unless otherwise noted, the SST was
299K as in Bryan and Rotunno (2009), the lower SST being motivated by this sounding’s cooler
surface air temperature (see also Bryan 2012). All CM1 fields shown are averaged between days 10 through 12, inclusive, except in Sec. 3d, in which an 8–12 day averaging interval was adopted for consistency with Bryan and Rotunno (2009).

3. Results

a. PBL cooperation with CRF

As reviewed above, Bu et al. (2014) demonstrated that CRF plays an important role in determining TC structure. This can be seen in HWRF simulations made using Thompson (“T”) microphysics and the GFS PBL scheme with either RRTMG (labeled CRF-on) or GFDL radiation (labeled CRF-off) for $\alpha = 0.7$ and 0.25 in (1), representing the 2014 operational model setting and the recommendation of Gopalakrishnan et al. (2013), respectively. Enabling CRF can increase the storm size (as manifested by the 10-m $R_{34}$) by a substantial (and MP-dependent) amount (compare solid and dashed red, or solid and dashed blue contours in Fig. 1) because hydrometers interact with radiation to force gentle ascent, elevating the relative humidity through a deep layer mainly above the PBL, resulting in enhanced convective activity in the TC outer core. Although some details (including magnitude) are dependent on microphysics, resolution, and other factors, the expected pattern of net cooling along cloud top with warming through much of the cloudy area is seen in the T/RRTMG simulation (Fig. 2a), but not in its T/GFDL counterpart (Fig. 2b). The anvil cloud in the case with effective CRF is thicker and also wider, in part because cloud-radiative forcing itself acts as a positive feedback on anvil extent, as demonstrated in Fovell et al. (2016) (see their Fig. 11.20).

After we identified the GFDL scheme’s lack of cloud-radiative forcing for deep clouds, the Developmental Testbed Center (DTC) and the HWRF team evaluated the RRTMG scheme along
with the Thompson MP for adoption in the operational HWRF (see Introduction). Their analyses of retrospective simulations demonstrated that the HWRF forecast skill was generally degraded when the new physics was included and, as a consequence, neither package was adopted for the 2014 TC season. The T/RRTMG model storms developed a positive size bias, which was especially pronounced among the Atlantic cases. In the East Pacific subset, the T/RRTMG cases tended to exhibit positive biases early on, but encountered colder SSTs sooner, resulting in negative size biases at longer forecast lead times.

Our working hypothesis was that excessive mixing associated with the GFS scheme with $\alpha \approx 1$, including the value selected for the operational model, compensated for the model’s tendency to produce overly small TCs, *which was actually a consequence of the missing CRF*. Therefore, when the radiation problem was fixed, the model was left with a positive size bias. Put another way, we believe that an improved implementation of CRF justifies a smaller value of vertical eddy diffusivity, at least in the context of the GFS PBL scheme$^2$.

Consistent with this interpretation, Fig. 1 reveals that CRF and $\alpha$ have qualitatively similar influences on horizontal storm size. Note that varying $\alpha$ (for fixed CRF) causes the 34-kt wind radius to increase significantly independent of the radiation scheme employed (compare red and blue contour pairs). For instance, with GFDL radiation, increasing $\alpha$ from 0.25 (blue dashed) to 0.7 (red dashed) shifts $R_{34}$ from 90 to 150 km. The narrowest storm used the $\alpha = 0.25$ value suggested by Gopalakrishnan et al. (2013) with GFDL radiation, while the widest employed the 2014–15 operational setting (0.7) with RRTMG. Thus, it is seen that the physics interplay between CRF and mixing can alter the 34-kt wind radius by factor of two, and there is a material impact on the eye size as well.

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$^2$It is noted that for the 2015 HWRF (Tallapragada et al. 2015), the $\alpha$ parameter was replaced by a strategy that does not require an externally-set free parameter. See Bu and Fovell (2015) for more information.
We note that the range of $R_{34}$ found in the experiments above, e.g., 90–205 km, is consistent with observations of TC size derived from ships, buoys, aircraft reconnaissance, and satellite-derived algorithms (see review in Knaff et al. 2016). These include an interquartile range of 138–277 km in the Atlantic basin extended best track (Kimball and Mulekar 2004), 1.8° mean (1° standard deviation) satellite-derived 34-kt radius of Wu et al. (2015) in western North Pacific TCs, and range of 90–300 km in Atlantic basin storms with concurrent Hurricane Wind Analysis System (H*Wind) and QuikSCAT data (Chavas et al. 2015).

b. Vertical eddy mixing influence on storm size

1) Sensitivity to $\alpha$

The expansion of the 34-kt wind radius seen in Fig. 1 occurs because the PBL mixing acts in a very similar manner as CRF in expanding storm size, as illustrated by the temporally- and spatially-averaged microphysical diabatic tendency and tangential wind fields shown in Fig. 3. Implementing CRF for fixed $\alpha$ (right column) and varying $\alpha$ with CRF active (left column) both result in a radially more extended heating field, causing the wind field (as illustrated by the tangential wind differences in the bottom row) to expand outward in qualitatively similar manners, for the reasons discussed in Bu et al. (2014). The impact on the eye size is also obvious in the difference plots. Note that while the GFDL/$\alpha = 0.7$ and RRTMG/$\alpha = 0.25$ runs possessed nearly identical 10-m wind profiles (Fig. 1) their tangential winds differed more substantially aloft. We are focusing on the 10-m winds because these are used in skill assessments. However, these results serve as a reminder that the near-surface winds alone may not be sufficient to accurately determine actual storm size.

As noted above, the $\alpha$ parameter was added to (1) to control eddy mixing in the TC inner core with the GFS PBL scheme. Figure 4a shows vertical profiles of $K_m$ from T/RRTMG/GFS
simulations, averaged over an annulus residing between 30 and 200 km from the center for various $\alpha$ values, with the corresponding near-surface wind profiles presented in Fig. 5a. The profiles differ little with respect to vertical shape, which is determined by (1) and the PBL depth $h$. However, increasing $\alpha$ causes $R_{34}$ to increase monotonically from 150 km to more than 250 km, and for the RMW to shift outward as well. These profiles represent very different storm sizes but essentially the same fundamental structures when nondimensionalized with respect to maximum wind speed (Vmax) and RMW (Fig. 5b). The modified Rankine slope parameter between one and four multiples of the RMW is about 0.65, very near the upper bound determined by Mallen et al. (2005) for all TCs ranging from pre-hurricane to major hurricane intensity.

One important and direct impact of the eddy mixing is associated with the vertical transport of water vapor in the boundary layer, upward from the sea surface to the PBL top. Figure 6’s left panels, which present water vapor and $K_h$ fields for T/RRTMG storms with $\alpha=0.7$ and 0.25 along with their differences, demonstrate that the more substantial mixing produced with larger $\alpha$ is associated with higher moisture content in the upper portion of the PBL, especially at larger radii. This pattern is consistent with the contribution of vertical eddy mixing to the local water vapor ($q_v$) tendency, which is a second-order parabolic term of the form

$$\left[ \frac{\partial q_v}{\partial t} \right]_{mix} = \frac{\partial}{\partial z} \left( K_h \frac{\partial q_v}{\partial z} \right).$$

In the atmosphere, the water vapor concentration decreases quasi-linearly with height and, as a consequence of the parabolic vertical shape of $K_h$, we would expect negative vapor tendencies where $K_h$ increases with height (below the level where $K_h$ reaches its maximum) and positive tendencies where $K_h$ decreases with height (above the $K_h$ maximum). This also applies to the difference fields, and explains the positive values above, and negative ones below, the level of maximum $K_h$ difference (Fig. 6c). Thus, one contributor leading to the greater PBL moisture ($\geq$ ...
400 m above the sea surface) in the larger $\alpha$ run is enhanced vertical mixing. The enhanced water vapor transport to the top of the PBL brings the air there closer to saturation, which can encourage more convective activity, producing the diabatic heating that eventually leads to a broader wind field (see discussion in Bu et al. 2014 and Fovell et al. 2016).

Figure 6’s right panels extend the comparison to two simulations varying CRF for fixed $\alpha$ and illustrate two important points. First, CRF itself induces a change in the PBL mixing. This is not surprising as the parameterized mixing responds to the circulation changes induced by cloud-radiation interaction. Second, the effect of altered mixing on the vapor field in this experiment is dominated by the CRF influence, which is sizable and not confined to the boundary layer. As a consequence, we will explicitly control the PBL mixing in some subsequent sensitivity tests in order to separate these two effects.

2) INFLUENCE OF SST ON $\alpha$ SENSITIVITY

Examination of DTC’s HWRF retrospective cases from their initial Thompson and RRTMG tests described above suggested to us that the impact of $\alpha$ could vary from case to case, and even from region to region, with some TCs being quite insensitive to the value employed. From these cases, we surmised that the less convectively favorable the environment, the less influence eddy mixing of moisture would, or could, have. Within the semi-idealized framework, we can establish a more unfavorable environment by simply lowering the SST from its standard value of 302.5 K. In this subsection, we explore how SST modulates the impact of $\alpha$ on the storm size, selecting values of 300 and 298 K to examine.

Colder SSTs result in smaller and weaker storms, other factors being equal (Fig. 7), consistent with Holland (1997), Lin et al. (2015), and Chavas et al. (2016), and the disparity between the larger and smaller $\alpha$ diminishes as well. While the water vapor and $K_h$ difference fields for
these cooler SST cases (Fig. 8) resemble those of the standard case (Fig. 6c), the magnitudes are
markedly smaller. Less vapor is being transported upward through the PBL and, as a consequence,
outer core convective activity and winds are correspondingly weaker. Thus, the storms are more
compact.

In this experiment, there are two convolved factors: the diminished entropy supply from the sea
surface directly reduces the storm intensity, but also indirectly decreases the eddy mixing since $K_m$
(and $K_h$) is proportional to $U_*$, which itself depends on the near-surface wind speed. These factors
can be separated in a straightforward way with an axisymmetric version of the CM1 model, which
uses a version of Thompson microphysics and a radiation scheme (Goddard) that is comparable
to RRTMG. In this special assessment, CM1’s vertical mixing is deactivated and replaced with
azimuthally- and temporally-averaged $K_m$ and $K_h$ fields derived from the HWRF’s control run
with $\alpha=1$ that, although not explicitly shown, closely resembles that in Fig. 6a but with larger
magnitude (see also Fig. 4). We set $K_m = K_h$ so $Pr = 1$ as is typical in the GFS scheme. These
fields are imposed from the initial time and held fixed, making them independent of the model
physics and dynamics, and their magnitudes easy to manipulate.

Figure 9 shows temporally-averaged 10-m wind profiles for four different SSTs for these “fixed-
$K$” simulations using both the original and decreased amounts of mixing. For the latter, $K_m$ and $K_h$
are still equivalent but reduced by a factor of 3. It is worth noting that these CM1 storms tend to be
stronger (in terms of maximum 10-m wind) than their HWRF counterparts, in part because they
cannot develop asymmetries. However, it is clear that as the SST is lowered, the storms become
both smaller and less sensitive to the magnitude of the eddy mixing, the radial size differences
being 42%, 31%, and 21% for SSTs of 299 K, 297 K, and 295 K, respectively. Less favorable
environments contain less available vapor and thus mixing has a diminished influence on storm
structure.
Thus, it appears that TC size can be directly modulated via water vapor transport in the boundary layer, with the sensitivity to $\alpha$ values gradually disappearing as the entropy supply from the sea surface declines. This reveals that the inclusion of lower sensitivity cases could serve to partially obscure the influence of PBL mixing in ensemble statistics incorporating a large number of events.

We note in passing that in both the HWRF and CM1 mixing experiments (Figs. 7 and 9), the intensity difference between greater and lesser mixing is not a simple function of SST. TC intensity is a complex function of many factors, including available energy from the sea-surface as well as the competition between inner- and outer-core convective activity (e.g., May and Holland 1999; Wang 2009; DeMaria et al. 2012). Although decreasing the water vapor diffusion through the whole domain may suppress the convection in the eyewall region somewhat, the outer convection may be reduced even more. As a consequence, the net influence may be to actually intensify the TC; this deserves further study.

3) INFLUENCE OF $\alpha$ ON SCALARS AND MOMENTUM

We have shown that the unmodified GFS PBL parameterization produces vigorous mixing and reducing this via $\alpha$ results in the model storms becoming smaller. Since $K_h = K_m Pr^{-1}$ and the scheme yields $Pr \approx 1$ in the hurricane boundary layer, this means that manipulating $\alpha$ modifies the momentum and scalar mixing equally. We now examine HWRF simulations in which $K_m$ or $K_h$ are reduced separately, without explicitly modifying the other mixing coefficient, using the $\alpha = 1$ simulation included in Fig. 5a as the control run. This can be considered a selective application of $\alpha$ and/or a direct manipulation of $Pr$, the handling of which varies among parameterizations. As an example, the YSU scheme, the subject of the next subsection, develops $Pr < 1$ in the hurricane PBL, meaning relatively larger diffusion is applied to scalars than to momentum.
The experiment reveals that the mixing applied to scalars (being water vapor, temperature, and non-precipitating condensate for GFS) has a greater influence on storm size than that applied to momentum. Reducing the mixing applied to these fields by two-thirds (without explicitly modifying $K_m$) results in a substantially narrower storm than in the control run, not only at the 10-m level (Fig. 10a) but also through an appreciably deep layer, as revealed by vertical profiles of wind speed averaged through the 100–250 km annulus (Fig. 10b) and vertical cross-sections of tangential wind and diabatic heating (Fig. 11a,b). Consistent with our prior findings, the diabatic heating fields suggest the narrowness is a consequence of diminished convective activity beyond the TC inner core. Furthermore, this result is mainly driven by vapor diffusion, as reducing the mixing applied to water vapor alone suffices to accomplish most of the storm contraction [simulation labeled $K_h/3$ (vapor only)].

In contrast, manipulating the momentum vertical diffusion (without explicitly modifying $K_h$) has a much smaller overall impact on storm width. In this example, the effect appears large (and comparable to scalar mixing reduction) at the 10-m level (Figs. 10a,b), but what has changed most is the near-surface vertical shear as the wind profile farther aloft is less affected (Figs. 10b, 11c). This result may be anticipated with a version of (2) acting on the horizontal wind speed $U$:

$$\left[ \frac{\partial U}{\partial t} \right]_{mix} = \frac{\partial K_m}{\partial z} \frac{\partial U}{\partial z} + K_m \frac{\partial^2 U}{\partial z^2}. \quad (3)$$

The first term on the right hand side dominates because the wind profile just above the surface lacks curvature. $K_m$ increases with height to $Z \approx 0.5$ km (Fig. 4a), so the wind speed tendency due to mixing below that level is positive, representing the transport of higher wind speeds downward towards the surface and reducing the vertical shear established by friction. Therefore, when the magnitude of $K_m$ is directly restrained, as in the $K_m/3$ experiment, the wind speed at the 10-m level
is impacted, although the influence farther aloft is smaller, increasing the vertical shear. This is another illustration that wind information from a single level may be deceptive.

An alternative explanation of this result may also be pursued by expressing (3) as

$$\left[ \frac{\partial U}{\partial t} \right]_{mix} = \frac{\partial}{\partial z} \left( K_m \frac{\partial U}{\partial z} \right).$$  \hspace{2cm} (4)

At the lowest model level, the vertical gradient appears to involve separate $K_m$ values provided by (1) and implied by the surface layer parameterization, which under standard assumptions (cf. Stensrud 2007; Kepert 2012) is

$$K_m = \kappa \left( \frac{U_*}{\phi_m} \right) Z.$$  \hspace{2cm} (5)

These assumptions also lead to the logarithmic wind profile under neutral conditions. Augmenting (1) with $\alpha$ means it no longer approaches (5) as $Z \to 0$. As a consequence, selecting $\alpha < 1$ artificially increases the vertical gradient of $K_m$, other factors being equal. This has the effect of strengthening the frictional drag and, thus, enhancing the wind shear near the lower boundary.

Whatever the explanation, it is clear that though the vertical shear clearly varies inversely with $\alpha$, the normalized wind profiles from the GFS $\alpha < 1$ experiments remain logarithmic near the surface (Fig. 12). Also shown are the $K_m/3$ and $K_h/3$ (orange dotted and dashed, respectively) cases as well as runs made using the YSU (black; see also Sec. 3c), MYJ (grey dotted), and QNSE\(^3\) (grey dashed) PBL schemes. Note that the shear in the $\alpha = 1$ storm is larger than that produced by QNSE and YSU but comparable to the MYJ case, and that manipulating $K_h$ independently of $K_m$ has little impact on the shear, which was also suggested by Fig. 10b. The important point is that while the $\alpha$ parameter modulates the mixing of both momentum and scalars, it is the moisture diffusion within the mixed layer that most strongly modulates storm size through the troposphere in our experiment.

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\(^3\)Quasi-Normal Scale Elimination; Sukoriansky et al. 2006.
Since modifying the mixing applied to some fields can alter the entire circulation, the diffusion not being directly manipulated is also affected to some degree. As a consequence, we also consider a version of the last subsection’s fixed-\( K \) CM1 experiment in which \( K_m \) and \( K_h \) are still externally imposed but now manipulated separately, again by reducing the coefficients by two-thirds (also shown on Fig. 9, as dashed curves). We again find that \( K_m \) has little influence on the storm size as determined from the 10-m wind profile, while reducing \( K_h \) results in narrower storms with smaller eyes, with the differences being more substantial in more favorable environments. The storms with reduced scalar mixing are also stronger than their reduced momentum mixing counterparts, with the apparent exception of the SST = 300 K case. That particular model storm develops an eyewall replacement cycle during the averaging period (not shown), leading to a lower mean intensity.

c. Comparison with YSU scheme

Since its inception, the HWRF model has used some version of the GFS PBL scheme, while the YSU parameterization is a popular choice with the Advanced Research WRF (ARW) core. As these non-local schemes evolved from a common ancestor, they unsurprisingly retain many similarities, including the same prescribed parabolic shape function for eddy mixing below the boundary layer depth, \( h \). In our tests, however, YSU tends to produce shallower boundary layers that, owing to (1), make the eddy mixing magnitudes smaller and shift the level of maximum mixing closer to the surface (solid black curve on Fig. 4b). One can now anticipate this has an impact on vertical moisture transport and, thus, storm radial extent. At 10-m MSL, the YSU storm’s \( R_{34} \) is about 182 km, comparable to the GFS simulation with \( \alpha = 0.4 \) (Fig. 5c).

There are several potentially influential differences between the current YSU and GFS implementations, including their handling of the turbulent Prandtl number (as noted earlier) and the free atmosphere above the PBL, as well as the present need to employ different surface layer
parameterizations. However, in the present study, by far the most important factor involves the specification of the critical bulk Richardson number, $Rib_{cr}$, which influences the PBL height, $h$, with larger $Rib_{cr}$ resulting in greater boundary layer depths. In the original MRF scheme (Hong and Pan 1996), the $Rib_{cr}$ value of 0.5 suggested by Troen and Mahrt (1986) was adopted, while recent practice with the GFS scheme in HWRF has been to set $Rib_{cr} = 0.25$ over water, with optional modification based on the surface Rossby number (cf. Vickers and Mahrt 2004). Although originally set to 0.5 (Hong et al. 2004), YSU evolved to employ different $Rib_{cr}$ values for unstable and stable conditions, and currently uses $Rib_{cr} = 0.0$ for the unstable PBL (Hong 2010).

YSU’s stability-dependent handling of $Rib_{cr}$ results in the relatively shallow boundary layer depth seen in Fig. 4b. When the default YSU scheme is altered to adopt a $Rib_{cr}$ of 0.25, as in the current GFS parameterization, the fields and storm structures more closely resemble the GFS results seen earlier (Fig. 4b, 5c). There is more substantial mixing over a greater depth (compare Figs. 13a and b), producing a larger vertical transport of water vapor above about 700 m MSL (Fig. 13c)$^4$. Averaged through the 30–200 km annulus, the modified YSU mixing (dashed black curve on Fig. 4b) closely resembles that produced by the GFS scheme when $\alpha = 0.7$. The wind field at 10-m MSL is also expanded somewhat (Fig. 5c), and comparable to GFS with $\alpha = 0.7$.

As the GFS and YSU parameterizations possess other differences that can impact the hurricane circulation, we consider yet another fixed-$K$ experiment in which the effect of mixing depth is explored (Fig. 14) in a more controlled fashion. For this experiment, the GFS-supplied $K_m$ and $K_h$ fields are either jointly or separately contracted vertically by 50%, mimicking the PBL depths YSU produces in HWRF, but without change in magnitude. Relative to the control configuration (solid black curve), halving the depth of both eddy mixing fields (solid red curve) results in a storm that

$^4$This comparison involves $K_h$ as the YSU scheme permits $Pr$ to become smaller than unity, resulting in larger mixing being applied to scalars than to momentum.
is narrower in every respect as well as stronger in intensity. As anticipated from prior results, this radial contraction is associated with reduced convective activity in the outer region (not shown). The intensification and width reduction is a consequence of the alteration of the scalar mixing alone (dashed red curve), as halving the depth of just $K_m$ actually results in a small increase of storm width, other factors being equal.

Up to this point, we have focused on azimuthally-averaged fields, partly for simplicity. However, this disguises the differences among the storms with respect to asymmetric structures, especially beyond the TC inner core. Figure 15 presents mass-weighted mean vertical velocity between the surface and 500 hPa from the HWRF simulations, again temporally averaged over the simulations’ final diurnal cycle. The YSU storm (Fig. 15a) is compact, with a narrow and asymmetric eye, and relatively little outer rainband activity. Raising $Rib_{cr}$ to 0.25 (Fig. 15b) results in an enhanced primary rainband structure (cf. Houze 2010), more closely resembling what the GFS scheme produces with $\alpha = 0.4$ (Fig. 15c). This feature is most prominent when the GFS eddy mixing is even less constrained (Fig. 15d). As these are semi-idealized experiments, there is no correct answer, but it remains that PBL mixing is clearly influential in modulating outer storm structure.

**d. Comparison with selected axisymmetric studies**

In contrast to our study, the axisymmetric studies of Bryan and Rotunno (2009), Bryan (2012), Chavas and Emanuel (2014), and Frisius (2015) reported little sensitivity of TC size to vertical mixing, which can be manipulated via the vertical mixing length, $l_v$. Bryan (2012) found “a slight tendency for smaller $R_{34}$ as $l_v$ decreases”, and variations of 20–28 km are seen for that paper’s two setups when his recommended values for horizontal mixing length ($l_h = 1000$ m) and the enthalpy/drag coefficient ratio ($C_k/C_d = 0.5$) were adopted (see his Fig. 7). Frisius (2015) suggested that the lifetime of a TC is too short for vertical mixing sensitivity to be relevant.
However, there are a number of potentially important differences between their experiments and ours, the most influential one being the treatment of atmospheric radiation.

In this subsection, we employ axisymmetric CM1 simulations configured similarly to Bryan and Rotunno (2009), albeit with somewhat coarser (3 km) radial grid spacing and Thompson microphysics. Vertical mixing lengths of 100 and 25 m are examined with \( l_h \) fixed at 1000 m. The cited axisymmetric studies employed somewhat different, and yet all highly simplified, radiation treatments in lieu of a full parameterization, so we adopt the approach employed by Bryan and Rotunno (2009) in which a sponge term mimicking clear-sky cooling is added to the temperature equation. This approach, which naturally lacks a diurnal cycle and ignores CRF, is identified as “R-E relaxation” after Rotunno and Emanuel (1987). As in Bryan and Rotunno (2009), cooling is capped at 2 K day\(^{-1}\). The \( C_k/C_d \) ratio varies with wind speed, but remains in the range from 0.5–0.7 between the RMW and \( R_{34} \).

Owing to friction, a storm’s fastest winds are typically located hundreds of meters above the surface, and both R-E relaxation cases attain temporally-averaged peak tangential velocities [computed in the manner of Bryan and Rotunno (2009)] of about 90 m s\(^{-1}\) (not shown), consistent with the values provided in Fig. 2 of Bryan and Rotunno (2009). At 10 m MSL, however, both are quite compact relative to the HWRF simulations examined previously, with \( R_{34} \) of only about 65 km (blue curves in Fig. 16a). The narrowness persists through the 12-day simulation period following maturity (Fig. 17a) and extends through the troposphere, associated with a very clear absence of outer convective activity (shown for \( l_v = 100 \) m in Fig. 18a). The vertical eddy mixing field (shown for \( l_v = 100 \) m in Fig. 19b) and applied equally to momentum and scalars bears some resemblance to the HWRF runs using GFS (Fig. 6a) in that both have a parabolic vertical shape with a maximum at about 0.5–0.6 km MSL beyond the RMW. While fairly large values of mixing
extend vertically into the eyewall, a characteristic explained by Kepert (2012), note the mixing
strength tapers off much more rapidly in the radial direction.

In contrast, the CM1 simulations employing Goddard radiation (with CRF active) yield much
wider storms (black curves in Fig. 16a), which are more comparable to the HWRF simulations and
expand progressively with time (Fig. 17b). These results are consistent with the findings of Hakim
(2011). There is a suggestion of more outer convective activity (shown for $l_v = 100$ m in Fig. 18b)
and greater sensitivity to the vertical mixing length (Figs. 16a, 17b). In line with the broader
and stronger storm circulation, the vertical eddy mixing field is larger in both magnitude and
radial extent (Fig. 19a). As anticipated from prior results, the more vigorous mixing contributes to
increasing the water vapor content in the upper portion of the boundary layer relative to the R-E
case (Fig. 19c).

The three principal differences between the R-E relaxation and Goddard/CRF runs for a given
$l_v$ setting are the manner in which clear-sky radiative cooling is computed, inclusion or exclusion
of cloud-radiative feedback, and the amount of the boundary layer eddy mixing. Our analysis
suggests all three factors contribute to making the Goddard/CRF storm wider, with eddy mixing
being the least important. First, the effect of the handling of clear-sky atmospheric radiation
alone is tested using Goddard radiation but with CRF deactivated. Relative to the R-E simulation,
the CRF-off storm is both stronger and wider (black curve in Fig. 16b; Fig. 18c), although still
narrower than its CRF-active counterpart and with a much slower rate of radial expansion. Even
with transparent clouds, however, organization is more rapid and greater sensitivity to $l_v$ is apparent
(Fig. 17c).

Second, the effect of eddy mixing in isolation can be tested via a final fixed-$K$ experiment,
this time utilizing the R-E case’s temporally averaged $K_m (= K_h)$ field (Fig. 19b). Again, this is
externally applied from the start of the simulation, and Goddard radiation is employed with CRF
active. Despite the relatively weak and restricted mixing at outer radii, the storm is still able to organize rapidly (Fig. 17d), develop radially extensive convective activity (Fig. 18d), and attain a 10-m wind profile comparable to the standard CRF example (red curve in Fig. 16b). That this is mostly due to CRF is demonstrated by the much slower expansion of this experiment’s CRF-off version (20 m s\(^{-1}\) contour superposed in green in Fig. 17d), which more closely resembles the other CRF-off storm (Fig. 17c) with respect to size and expansion rate. In agreement with our previous fixed-\(K\) experiments, these results suggest that a vertical mixing field that evolves with time to become radially extensive can assist in the progressive expansion of a tropical cyclone, but is not absolutely necessary, particularly when clouds are permitted to interact with radiation.

4. Discussion and summary

Bu et al. (2014) demonstrated that cloud-radiative forcing (CRF) can exert a substantial influence on numerically simulated tropical cyclones (TCs), especially with respect to the storm’s horizontal scale. Specifically, it is the within-cloud longwave warming component of CRF that indirectly enhances convective activity in the TC outer core, thereby generating the diabatic heating that broadens the wind field. They further established that the radiation scheme employed by the operational HWRF model, which derived from the old GFDL parameterization, was very deficient in handling CRF, to the point that it was essentially absent. However, as mentioned in the Introduction, when the HWRF model was applied to historical cases using a more realistic radiation package, model skill with respect to important storm characteristics, such as intensity, position, and size, was degraded. In particular, most storms developed a positive size bias with respect to \(R_{34}\), the radius of the 34-kt (17.5 m s\(^{-1}\)) wind at 10 m above the surface, one of the important metrics used in model verification.
That result motivated a study of how and why the planetary boundary layer (PBL) and its parameterization affect storm size, in cooperation and competition with CRF. Our principal finding is that vertical mixing of scalars in the boundary layer, primarily water vapor, also influences storm size via modulating outer core convective activity. In this case, it is eddy mixing that helps transport water vapor to the top of the boundary layer, elevating the relative humidity there, and making the outer core more favorable for convection. Thus, a compelling parallel is seen with respect to CRF, the difference being that the heating associated with in-cloud warming was focused above the boundary layer, while the PBL influence is essentially “bottom-up” from the sea-surface.

As a consequence, we see why TC structure is sensitive to the PBL parameterization, particularly with respect to the magnitude and shape of the eddy diffusion distributions they generate. This was demonstrated using HWRF’s operational boundary layer code, the GFS PBL scheme, which was modified in the past to incorporate a tuning parameter, $\alpha$, to throttle mixing in the TC inner core (Gopalakrishnan et al. 2013) because observations (e.g., Zhang et al. 2011) suggested the model was too diffusive. We showed that $\alpha$ has a profound influence on $R_{34}$. Indeed, we believe that it was the excessive diffusion produced by the GFS scheme that was previously compensating for the lack of cloud-radiative forcing in the HWRF model such that when the radiation issue was fixed, the simulated TCs developed a positive size bias, leading to poorer wind structure, intensity, and position forecasts.

Eddy mixing applied to momentum also appears to influence storm size, at least relatively close to the surface. However, what changed the most was the vertical shear, as the winds farther aloft was less impacted. Since the breadth of the 10-m MSL wind field is one of the parameters used to judge model forecast skill, and information from farther aloft is often absent, this result raises the possibility that available storm size and/or intensity information can be skewed or misinterpreted. That said, examination of retrospective cases made using the HWRF model suggested that TCs
are not always responsive to variations in the eddy mixing and the sensitivity is diminished when
the TC environment is generally less favorable, which we demonstrated via experiments in which
the sea-surface temperature (SST) was lowered.

The GFS and YSU PBL schemes share a common ancestor, and represent parameterizations
that determine PBL height based on near-surface vertical stability and wind shear (cf. Vickers and
Mahrt 2004). We demonstrated that how the critical Richardson number, $R_{ibcr}$, was specified in
these schemes had a major impact on the magnitude and depth of the eddy mixing, indirectly
influencing TC size through vertical diffusion of boundary layer scalars. This finding motivates
closer study in the future as it appears to provide a more physically defensible way of modulating
mixing, especially because there is uncertainty with respect to the structure of the hurricane
inner core and how the PBL depth should be defined (cf. Zhang et al. 2011). Finally, through
a comparison with recent axisymmetric studies, we appreciate that vertical mixing can indeed
influence the progressive expansion of a TC in this restricted physical framework, but it does so
most efficiently when acting with the assistance of cloud-radiative forcing.

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HWRF simulations with GFS PBL ($\alpha = 0.7$)

(a) T/RRTMG
(b) T/GFDL

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Fig. 2. Temporally- and azimuthally-averaged total condensation (shaded, note logarithmic scale) and net radiation (negative [dashed] and positive [solid] contour interval 0.1 K h$^{-1}$) for Thompson/GFS storms using $\alpha = 0.7$ with (a) RRTMG (labeled CRF-on), and (b) GFDL (labeled CRF-off) radiation.
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**HWRF simulations**

**Temporally, azimuthally, and radially averaged horizontal wind profiles**

![Graph](image)

**FIG. 12.** Vertical profiles of the horizontal wind, averaged in time and through an annulus extending from 100 to 250 km from the storm center, and normalized with respect to the wind speed at about 300 m MSL, from GFS runs with $\alpha = 1, 0.75, 0.5,$ and $0.25$; the $K_m/3$ and $K_v/3$ (all scalars) tests; and simulations with the YSU, MYJ, and QNSE PBL schemes. Black squares on the the GFS $\alpha=1$ profile indicate model levels. The $K_v/3$ (vapor only) case is indistinguishable from the $K_m/3$ (all scalars) profile shown.
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FIG. 15. Mass-weighted mean vertical velocity (m s$^{-1}$) from HWRF T/RRTMG simulations using (a) the default YSU scheme with Rib$_{cr}=0.0$, (b) YSU with Rib$_{cr}=0.25$, (c) the GFS scheme with (c) $\alpha=0.4$, and (d) GFS with $\alpha=0.7$. Range rings of 50 and 150 km depicted, and top of plots represents north.
FIG. 16. 10-m wind speed from CM1 simulations, averaged between days 8 and 12. (a) Simulations using R-E relaxation (blue curves) and Goddard radiation with CRF on (black curves) with $l_v = 100$ and 50 m (solid and dashed, respectively). (b) Simulations using Goddard radiation with CRF off and $l_v = 100$ m (black curve) and a fixed-$K$ run using the R-E ($l_v = 100$ m) simulation’s eddy mixing field (red curve). The $l_v = 100$ m profiles from (a) are included in grey for reference.
FIG. 17. Hovmöller (time vs. radius) diagrams of tangential wind speed at 10 m MSL, with 20 m/s contour bolded, for CM1 simulations with $l_v = 100$ m and using (a) R-E relaxation, (b) Goddard radiation with CRF on, (c) Goddard radiation with CRF off, and (d) Goddard with CRF on but with fixed eddy mixing from the R-E simulation in (a). On (a) and (b), 20 m/s contours from corresponding simulations using $l_v = 25$ m are shown in green. On (d), 20 m/s contours for CRF-off version shown in green.
FIG. 18. Radius vs. height cross-sections showing the temporally-averaged microphysics diabatic forcing (shaded) and tangential wind (m s$^{-1}$) from CM1 simulations using (a) R-E relaxation, (b) Goddard radiation with CRF on, (c) Goddard radiation with CRF off, and (d) Goddard with CRF on but with fixed eddy mixing from the R-E simulation in (a). 20 m s$^{-1}$ contours bolded. All simulations used $l_v = 100$ m and were averaged between days 8–12.
Fig. 19. Similar to Fig. 6, but for CM1 simulations using (a) Goddard with CRF on and (b) R-E relaxation, with (c) difference fields. Both simulations used $l_c = 100$ m and were averaged between days 8–12.