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

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ABSTRACT

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

25 1. Introduction

26 Tropical cyclone (TC) size is an important forecast metric as it directly and indirectly influences
27 TC motion, intensity, track, and storm surge (e.g., Fiorino and Elsberry 1989; Fovell and Su 2007;
28 Lin and Chavas 2012; Carrasco et al. 2014). There are a variety of metrics used to define the TC
29 size, including the radius of the outermost closed isobar, the radius of vanishing wind, and the
30 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
31 level (MSL) is used to define the storm size or width.

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

42 Bu et al. (2014) also demonstrated that the GFDL-derived radiation scheme that was long
43 employed operationally in the Hurricane Weather Research and Forecasting (HWRF) model
44 (cf. Tallapragada et al. 2014) did not handle CRF properly, resulting in deep clouds that were
45 effectively transparent. Testing revealed, however, that implementing an ostensibly superior
46 radiation scheme *degraded* model skill (L. Bernardet, M. Biswas, and C. Holt, personal
47 communications, 2014). Analysis of those results led us to consider how the planetary boundary

48 layer (PBL) influences storm size, in cooperation and competition with CRF, which is the subject
49 of this study.

50 It is widely appreciated that boundary layer processes play an important role in TCs (e.g., Smith
51 1968; Ooyama 1969; Emanuel 1986; Van Sang et al. 2008). Among these processes are mixing
52 acting on momentum and scalars such as temperature and moisture, the subgrid portion of which
53 can be represented via diffusion coefficients K_m and K_h , respectively. Models such as HWRF
54 typically handle vertical mixing within the PBL via a parameterization that presumes a local
55 or non-local closure (cf. Stensrud 2007; Kepert 2012) such as the Mellor-Yamada-Janjić (MYJ;
56 Janjić 1990) and Yonsei University (YSU; Hong et al. 2006; Hong 2010) schemes, respectively;
57 Nolan et al. (2009b) provide a succinct comparison of these two approaches. Eddy mixing in
58 local schemes depends on the specification of a vertical mixing length (l_v), stability, and either the
59 vertical wind shear or prognosed turbulent kinetic energy. The HWRF model (see Tallapragada
60 et al. 2014, 2015) has been using a non-local scheme called GFS (Global Forecast System) that,
61 like YSU, is based on Troen and Mahrt (1986) and evolved from the Medium-Range Forecast
62 (MRF) model's PBL parameterization (Hong and Pan 1996).

63 There have been many papers addressing the sensitivity of simulated TCs to PBL
64 parameterizations and assumptions (e.g., Braun and Tao 2000; Hill and Lackmann 2009; Nolan
65 et al. 2009a,b; Smith and Thomsen 2010; Kepert 2012). Most previous studies of the PBL–TC
66 relationship have focused on TC intensity, inner core convection, and/or the TC PBL structure. A
67 few studies, however, have explicitly examined the influence of the vertical diffusion on the storm
68 structure, most of them reporting no significant influence. For example, Bryan (2012) found the
69 34-kt wind radius to be only weakly sensitive (and the radius of maximum wind, or RMW, to
70 be insensitive) to l_v , at least when reasonable values of the horizontal mixing length (l_h) and the
71 enthalpy/drag coefficient ratio ($C_k/C_d = 0.5$) are used. Bryan and Rotunno (2009) and Chavas and

72 Emanuel (2014) both demonstrated that the RMW remained essentially unchanged with doubled
 73 l_v . While Frisius (2015) did find some sensitivity to vertical diffusion, he pointed out that the
 74 lifetime of real TCs is too short for the effect of vertical diffusion to become relevant.

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

$$K_m = \kappa(U_*/\phi_m)Z[\alpha(1 - Z/h)^2], \quad (1)$$

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

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

¹ Additionally, 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_h . This results in slightly different mixing being applied to moisture and heat.

(1). 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 $10\text{--}60\text{ m s}^{-1}$, 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 α 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

139 surface air temperature (see also Bryan 2012). All CM1 fields shown are averaged between days
140 10 through 12, inclusive, except in Sec. 3d, in which an 8–12 day averaging interval was adopted
141 for consistency with Bryan and Rotunno (2009).

142 **3. Results**

143 *a. PBL cooperation with CRF*

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

159 After we identified the GFDL scheme’s lack of cloud-radiative forcing for deep clouds, the
160 Developmental Testbed Center (DTC) and the HWRF team evaluated the RRTMG scheme along

161 with the Thompson MP for adoption in the operational HWRF (see Introduction). Their analyses
162 of retrospective simulations demonstrated that the HWRF forecast skill was generally degraded
163 when the new physics was included and, as a consequence, neither package was adopted for
164 the 2014 TC season. The T/RRTMG model storms developed a positive size bias, which was
165 especially pronounced among the Atlantic cases. In the East Pacific subset, the T/RRTMG cases
166 tended to exhibit positive biases early on, but encountered colder SSTs sooner, resulting in negative
167 size biases at longer forecast lead times.

168 Our working hypothesis was that excessive mixing associated with the GFS scheme with $\alpha \approx 1$,
169 including the value selected for the operational model, compensated for the model's tendency
170 to produce overly small TCs, *which was actually a consequence of the missing CRF*. Therefore,
171 when the radiation problem was fixed, the model was left with a positive size bias. Put another
172 way, we believe that an improved implementation of CRF justifies a smaller value of vertical eddy
173 diffusivity, at least in the context of the GFS PBL scheme².

174 Consistent with this interpretation, Fig. 1 reveals that CRF and α have qualitatively similar
175 influences on horizontal storm size. Note that varying α (for fixed CRF) causes the 34-kt wind
176 radius to increase significantly independent of the radiation scheme employed (compare red and
177 blue contour pairs). For instance, with GFDL radiation, increasing α from 0.25 (blue dashed)
178 to 0.7 (red dashed) shifts R_{34} from 90 to 150 km. The narrowest storm used the $\alpha = 0.25$ value
179 suggested by Gopalakrishnan et al. (2013) with GFDL radiation, while the widest employed the
180 2014–15 operational setting (0.7) with RRTMG. Thus, it is seen that the physics interplay between
181 CRF and mixing can alter the 34-kt wind radius by factor of two, and there is a material impact on
182 the eye size as well.

²It is noted that for the 2015 HWRF (Tallapragada et al. 2015), the α parameter was replaced by a strategy that does not require an externally-set free parameter. See Bu and Fovell (2015) for more information.

183 We note that the range of R_{34} found in the experiments above, e.g., 90–205 km, is consistent with
 184 observations of TC size derived from ships, buoys, aircraft reconnaissance, and satellite-derived
 185 algorithms (see review in Knaff et al. 2016). These include an interquartile range of 138–277 km
 186 in the Atlantic basin extended best track (Kimball and Mulekar 2004), 1.8° mean (1° standard
 187 deviation) satellite-derived 34-kt radius of Wu et al. (2015) in western North Pacific TCs, and
 188 range of 90–300 km in Atlantic basin storms with concurrent Hurricane Wind Analysis System
 189 (H*Wind) and QuikSCAT data (Chavas et al. 2015).

190 *b. Vertical eddy mixing influence on storm size*

191 1) SENSITIVITY TO α

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

204 As noted above, the α parameter was added to (1) to control eddy mixing in the TC inner
 205 core with the GFS PBL scheme. Figure 4a shows vertical profiles of K_m from T/RRTMG/GFS

206 simulations, averaged over an annulus residing between 30 and 200 km from the center for
 207 various α values, with the corresponding near-surface wind profiles presented in Fig. 5a. The
 208 profiles differ little with respect to vertical shape, which is determined by (1) and the PBL depth h .
 209 However, increasing α causes R_{34} to increase monotonically from 150 km to more than 250 km,
 210 and for the RMW to shift outward as well. These profiles represent very different storm sizes but
 211 essentially the same fundamental structures when nondimensionalized with respect to maximum
 212 wind speed (V_{max}) and RMW (Fig. 5b). The modified Rankine slope parameter between one and
 213 four multiples of the RMW is about 0.65, very near the upper bound determined by Mallen et al.
 214 (2005) for all TCs ranging from pre-hurricane to major hurricane intensity.

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

$$\left[\frac{\partial q_v}{\partial t} \right]_{mix} = \frac{\partial}{\partial z} \left(K_h \frac{\partial q_v}{\partial z} \right). \quad (2)$$

222 In the atmosphere, the water vapor concentration decreases quasi-linearly with height and, as
 223 a consequence of the parabolic vertical shape of K_h , we would expect negative vapor tendencies
 224 where K_h increases with height (below the level where K_h reaches its maximum) and positive
 225 tendencies where K_h decreases with height (above the K_h maximum). This also applies to the
 226 difference fields, and explains the positive values above, and negative ones below, the level of
 227 maximum K_h difference (Fig. 6c). Thus, one contributor leading to the greater PBL moisture ($\geq$

228 400 m above the sea surface) in the larger α run is enhanced vertical mixing. The enhanced water
229 vapor transport to the top of the PBL brings the air there closer to saturation, which can encourage
230 more convective activity, producing the diabatic heating that eventually leads to a broader wind
231 field (see discussion in Bu et al. 2014 and Fovell et al. 2016).

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

239 2) INFLUENCE OF SST ON α SENSITIVITY

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

248 Colder SSTs result in smaller and weaker storms, other factors being equal (Fig. 7), consistent
249 with Holland (1997), Lin et al. (2015), and Chavas et al. (2016), and the disparity between the
250 larger and smaller α diminishes as well. While the water vapor and K_h difference fields for

251 these cooler SST cases (Fig. 8) resemble those of the standard case (Fig. 6c), the magnitudes are
252 markedly smaller. Less vapor is being transported upward through the PBL and, as a consequence,
253 outer core convective activity and winds are correspondingly weaker. Thus, the storms are more
254 compact.

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

266 Figure 9 shows temporally-averaged 10-m wind profiles for four different SSTs for these “fixed-
267 K ” simulations using both the original and decreased amounts of mixing. For the latter, K_m and K_h
268 are still equivalent but reduced by a factor of 3. It is worth noting that these CM1 storms tend to be
269 stronger (in terms of maximum 10-m wind) than their HWRF counterparts, in part because they
270 cannot develop asymmetries. However, it is clear that as the SST is lowered, the storms become
271 both smaller *and* less sensitive to the magnitude of the eddy mixing, the radial size differences
272 being 42%, 31%, and 21% for SSTs of 299 K, 297 K, and 295 K, respectively. Less favorable
273 environments contain less available vapor and thus mixing has a diminished influence on storm
274 structure.

275 Thus, it appears that TC size can be directly modulated via water vapor transport in the boundary
276 layer, with the sensitivity to α values gradually disappearing as the entropy supply from the sea
277 surface declines. This reveals that the inclusion of lower sensitivity cases could serve to partially
278 obscure the influence of PBL mixing in ensemble statistics incorporating a large number of events.

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

287 3) INFLUENCE OF α ON SCALARS AND MOMENTUM

288 We have shown that the unmodified GFS PBL parameterization produces vigorous mixing and
289 reducing this via α results in the model storms becoming smaller. Since $K_h = K_m Pr^{-1}$ and the
290 scheme yields $Pr \approx 1$ in the hurricane boundary layer, this means that manipulating α modifies
291 the momentum and scalar mixing equally. We now examine HWRF simulations in which K_m or K_h
292 are reduced separately, without *explicitly* modifying the other mixing coefficient, using the $\alpha = 1$
293 simulation included in Fig. 5a as the control run. This can be considered a selective application of
294 α and/or a direct manipulation of Pr , the handling of which varies among parameterizations. As
295 an example, the YSU scheme, the subject of the next subsection, develops $Pr < 1$ in the hurricane
296 PBL, meaning relatively larger diffusion is applied to scalars than to momentum.

297 The experiment reveals that the mixing applied to scalars (being water vapor, temperature, and
 298 non-precipitating condensate for GFS) has a greater influence on storm size than that applied
 299 to momentum. Reducing the mixing applied to these fields by two-thirds (without explicitly
 300 modifying K_m) results in a substantially narrower storm than in the control run, not only at the
 301 10-m level (Fig. 10a) but also through an appreciably deep layer, as revealed by vertical profiles
 302 of wind speed averaged through the 100–250 km annulus (Fig. 10b) and vertical cross-sections of
 303 tangential wind and diabatic heating (Fig. 11a,b). Consistent with our prior findings, the diabatic
 304 heating fields suggest the narrowness is a consequence of diminished convective activity beyond
 305 the TC inner core. Furthermore, this result is mainly driven by vapor diffusion, as reducing
 306 the mixing applied to water vapor *alone* suffices to accomplish most of the storm contraction
 307 [simulation labeled $K_h/3$ (vapor only)].

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

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

313 The first term on the right hand side dominates because the wind profile just above the surface
 314 lacks curvature. K_m increases with height to $Z \approx 0.5$ km (Fig. 4a), so the wind speed tendency due
 315 to mixing below that level is positive, representing the transport of higher wind speeds downward
 316 towards the surface and reducing the vertical shear established by friction. Therefore, when the
 317 magnitude of K_m is directly restrained, as in the $K_m/3$ experiment, the wind speed at the 10-m level

318 is impacted, although the influence farther aloft is smaller, increasing the vertical shear. This is
 319 another illustration that wind information from a single level may be deceptive.

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

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

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

$$K_m = \kappa(U_*/\phi_m)Z. \quad (5)$$

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

328 Whatever the explanation, it is clear that though the vertical shear clearly varies inversely with α ,
 329 the normalized wind profiles from the GFS $\alpha < 1$ experiments remain logarithmic near the surface
 330 (Fig. 12). Also shown are the $K_m/3$ and $K_h/3$ (orange dotted and dashed, respectively) cases as
 331 well as runs made using the YSU (black; see also Sec. 3c), MYJ (grey dotted), and QNSE³ (grey
 332 dashed) PBL schemes. Note that the shear in the $\alpha = 1$ storm is larger than that produced by
 333 QNSE and YSU but comparable to the MYJ case, and that manipulating K_h independently of K_m
 334 has little impact on the shear, which was also suggested by Fig. 10b. The important point is that
 335 while the α parameter modulates the mixing of both momentum and scalars, it is the moisture
 336 diffusion within the mixed layer that most strongly modulates storm size through the troposphere
 337 in our experiment.

³Quasi-Normal Scale Elimination; Sukoriansky et al. 2006.

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

348 *c. Comparison with YSU scheme*

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

358 There are several potentially influential differences between the current YSU and GFS
 359 implementations, including their handling of the turbulent Prandtl number (as noted earlier) and
 360 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

⁴This 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.

406 However, there are a number of potentially important differences between their experiments and
407 ours, the most influential one being the treatment of atmospheric radiation.

408 In this subsection, we employ axisymmetric CM1 simulations configured similarly to Bryan
409 and Rotunno (2009), albeit with somewhat coarser (3 km) radial grid spacing and Thompson
410 microphysics. Vertical mixing lengths of 100 and 25 m are examined with l_h fixed at 1000 m. The
411 cited axisymmetric studies employed somewhat different, and yet all highly simplified, radiation
412 treatments in lieu of a full parameterization, so we adopt the approach employed by Bryan and
413 Rotunno (2009) in which a sponge term mimicking clear-sky cooling is added to the temperature
414 equation. This approach, which naturally lacks a diurnal cycle and ignores CRF, is identified as
415 “R-E relaxation” after Rotunno and Emanuel (1987). As in Bryan and Rotunno (2009), cooling
416 is capped at 2 K day^{-1} . The C_k/C_d ratio varies with wind speed, but remains in the range from
417 0.5–0.7 between the RMW and R_{34} .

418 Owing to friction, a storm’s fastest winds are typically located hundreds of meters above
419 the surface, and both R-E relaxation cases attain temporally-averaged peak tangential velocities
420 [computed in the manner of Bryan and Rotunno (2009)] of about 90 m s^{-1} (not shown), consistent
421 with the values provided in Fig. 2 of Bryan and Rotunno (2009). At 10 m MSL, however, both
422 are quite compact relative to the HWRF simulations examined previously, with R_{34} of only about
423 65 km (blue curves in Fig. 16a). The narrowness persists through the 12-day simulation period
424 following maturity (Fig. 17a) and extends through the troposphere, associated with a very clear
425 absence of outer convective activity (shown for $l_v = 100 \text{ m}$ in Fig. 18a). The vertical eddy mixing
426 field (shown for $l_v = 100 \text{ m}$ in Fig. 19b) and applied equally to momentum and scalars bears some
427 resemblance to the HWRF runs using GFS (Fig. 6a) in that both have a parabolic vertical shape
428 with a maximum at about 0.5–0.6 km MSL beyond the RMW. While fairly large values of mixing

429 extend vertically into the eyewall, a characteristic explained by Kepert (2012), note the mixing
430 strength tapers off much more rapidly in the radial direction.

431 In contrast, the CM1 simulations employing Goddard radiation (with CRF active) yield much
432 wider storms (black curves in Fig. 16a), which are more comparable to the HWRF simulations and
433 expand progressively with time (Fig. 17b). These results are consistent with the findings of Hakim
434 (2011). There is a suggestion of more outer convective activity (shown for $l_v = 100$ m in Fig. 18b)
435 and greater sensitivity to the vertical mixing length (Figs. 16a, 17b). In line with with the broader
436 and stronger storm circulation, the vertical eddy mixing field is larger in both magnitude and
437 radial extent (Fig. 19a). As anticipated from prior results, the more vigorous mixing contributes to
438 increasing the water vapor content in the upper portion of the boundary layer relative to the R-E
439 case (Fig. 19c).

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

450 Second, the effect of eddy mixing in isolation can be tested via a final fixed- K experiment,
451 this time utilizing the R-E case's temporally averaged K_m ($= K_h$) field (Fig. 19b). Again, this is
452 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.

475 That result motivated a study of how and why the planetary boundary layer (PBL) and its
476 parameterization affect storm size, in cooperation and competition with CRF. Our principal finding
477 is that vertical mixing of scalars in the boundary layer, primarily water vapor, also influences
478 storm size via modulating outer core convective activity. In this case, it is eddy mixing that helps
479 transport water vapor to the top of the boundary layer, elevating the relative humidity there, and
480 making the outer core more favorable for convection. Thus, a compelling parallel is seen with
481 respect to CRF, the difference being that the heating associated with in-cloud warming was focused
482 above the boundary layer, while the PBL influence is essentially “bottom-up” from the sea-surface.

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

493 Eddy mixing applied to momentum also appears to influence storm size, at least relatively close
494 to the surface. However, what changed the most was the *vertical shear*, as the winds farther aloft
495 was less impacted. Since the breadth of the 10-m MSL wind field is one of the parameters used to
496 judge model forecast skill, and information from farther aloft is often absent, this result raises the
497 possibility that available storm size and/or intensity information can be skewed or misinterpreted.
498 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, Rib_{cr} , 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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