**Impact of Dry Midlevel Air on the Tropical Cyclone Outer Circulation**

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**ABSTRACT**

The impact of dry midlevel air on the outer circulation of tropical cyclones is investigated in idealized simulations with and without a moist envelope protecting the inner core. It is found that a dry midlevel layer away from the cyclone center can broaden the outer primary circulation and thus the overall destructive potential at both developing and mature stages. The midlevel outer drying enhances the horizontal gradient of latent heating in the rainbands and drives the expansion of the outer circulation. The moist convection at large radii is suppressed rapidly after the midlevel air is dried in the outer rainbands. An enhanced horizontal gradient of latent heating initiates a radial-vertical overturning circulation anomaly in the rainbands. This anomalous overturning circulation accelerates the radial inflow of the main secondary circulation, increases the angular momentum import, and thus increases the cyclone size. The dry air, mixed into the boundary layer from the midtroposphere, is “recharged” by high enthalpy fluxes due to the increased thermodynamical disequilibrium above the sea surface. This “recharge” process protects the eyewall convection from the environmental dry air ventilation. The proposed mechanism may explain the continuous expansion in the tropical cyclone outer circulation after maturity as found in observations.

KEY WORDS: tropical cyclone; outer circulation; size; dry air; latent heating; midtroposphere

**1. Introduction**

It has long been recognized that the dry midlevel troposphere has an adverse effect on tropical cyclone (TC) genesis and intensification. Simpson and Riehl (1958) originally hypothesized that the environmental midlevel air with relatively low “heat content” (low-entropy or low-equivalent potential temperature) can be ventilated into the incipient vortex. This dry midlevel air from the environment may dilute the tropospheric heating in the inner core and discourage the TC development and maintenance (Gray 1968). Because of the importance of the initially central moist column to TC formation (Nolan 2007), the midlevel relative humidity (RH) in the environment has been taken as an imperative factor on understanding the temporally peaked distribution of TC genesis frequency during hurricane seasons in the tropical Atlantic (DeMaria et al. 2001), and predicting the incidence of cyclone genesis occurrences (e.g., Gray, 1975; Emanuel and Nolan, 2004).

The environmental humidity is important to TC intensity (e.g., Wu et al. 2015). By modeling a hurricane landfall event, Kimball (2006) showed that the environmental dry air was mainly wrapped inwards at 800-850 hPa. The environmental air sank into the boundary layer as it approached the core. The cyclone intensity was reduced when the dry air entrained into the eyewall updrafts. Instead of varying the humidity in the whole vertical air column (Kimball 2006), Braun et al*.* (2012) focused on the role of dry midlevel air in TC intensity. They found that the dry midlevel air has a deleterious impact on TC intensity but only if the dry air is close enough to the eyewall (e.g., no more than three times the initial radius of maximum wind). In addition to the magnitude of intensity change, the midlevel relative humidity near the cyclogenesis location is also found to play an important role in the lifetime maximum intensity (Ditchek et al. 2017) and the occurrence of rapid intensification at a later stage (Kaplan and DeMaria 2003).

Environmental humidity also plays an important role in TC outer circulation. Hill and Lackmann (2009) conducted a series of idealized numerical simulations by keeping an initial moist envelope surrounding the bogus vortex and reducing the relative humidity outside the radius of 100 km in the whole vertical air column. They found that the whole wind field expands in an initially moist environment. They explained the increase of TC size in an initially humid environment with the diabatic lateral expansion of the cyclonic potential vorticity. A negative effect of dry environment on TC outer circulation was also shown by Kimball (2006) in real case simulations. However, the response of TC outer circulation to dry midlevel air is somewhat ambiguous in Hill and Lackmann (2009) and not the focus of Kimball (2006) as both studies extended the moisture perturbations throughout the whole vertical column including the boundary layer.

Previous studies have mainly focused on the influence of environmental dry midlevel air on TC intensity. The impact of dry midlevel air on the TC outer circulation has only received a little attention. Braun et al. (2012) briefly mentioned the inhibiting effect of dry midlevel air on TC growth, if the dry air is inserted before the initial spinup and close to the core. Alland et al. (2017) specifically studied the effects of dry midlevel air on the development of TC secondary circulation with an axisymmetric model. They showed that the simulation with a moister midtroposphere has a wider overturning circulation. However, in their experiments, the dry midlevel air was also prescribed before the initial spinup and the TC core was exposed to the dry air directly during its development. To date, the impact of dry midlevel air on the outer circulation of an existing TC is still much less studied. In this study, we will show that the midlevel dry air from the ambient environment can, in fact, promote the expansion of the existing outer primary circulation.

We will investigate the impact of environmental dry midlevel air on TC outer primary circulation over an open warm ocean in a series of idealized numerical simulations. The environmental moisture will not be modified from the mean tropical sounding during the spinup stage, and it will be perturbed at both developing and mature stages by prescribing the midlevel RH.

The primary hypothesis is that after the initial spinup, the environmental dry midlevel air expands the outer primary circulation by forcing an enhanced radial-vertical overturning circulation in the rainbands, which is attributed to an enhanced horizontal gradient of latent heating due to the dry insertion in the midtroposphere. The next section introduces the model set-up and experiment design. Section 3 shows the simulation results. Section 4 discusses the main results, and the key findings are summarized in the last section.

**2. Methods**

**2.1 Model set-up**

The Weather Research and Forecasting (WRF) model (version 3.7.1; Skamarock et al*.*, 2008) is used to simulate TCs on an idealized aqua planet with a full-physics parameterization setup. The sea surface temperature (SST) is fixed as 27oC. TCs are simulated on an *f*-plane (20oN). Two square nested domains are two-way interactive. The side of the outer domain is 7200 km (600 × 600 grids with 12-km grid spacing) and the inner one is 3200 km (802 × 802 grids with 4-km grid spacing). The time steps for the outer and inner domains are 60 and 30 seconds, respectively. In each domain, there are 41 σ-levels and about 12 σ-levels located below 2-km height. The control simulation (CTRL) lasts for 9 days.

The initial bogus vortex is specified with an analytic wind profile model (Wang and Toumi 2016), which can be written as,

 (1)

where *Vo* is the near-surface wind profile of the bogus vortex, *r* is the radius from cyclone center, *Vmo*­ is the initial maximum wind speed that is set to 17 m s-1, *Rmo* is the initial radius of maximum wind speed that is set to 100 km, and *fo* is the Coriolis parameter that is set to 5×10-5 s-1. The wind speed decreases linearly to 0 m s-1 at the model top (20 km). The initial environmental condition in CTRL is specified with the tropical sounding during hurricane seasons (Jordan 1958). The background is set to be stationary. Thermal and dynamical fields are further adjusted after inserting the bogus vortex (Wang et al. 2015). The initial bogus vortex is placed in the center of the outer domain and the vortex-following technique is applied to the inner domain.

The lateral boundary conditions are fixed as the initial condition in the environment. A horizontal sponge layer with a width of 240 km is placed along the lateral boundaries of the outer domain at every vertical level. The sponge layer is designed to absorb noise along the boundaries by overwriting the horizontal wind velocities to zero at every time step. This sponge layer is important as it can effectively prevent the artificial boundary noise from influencing the TC wind field at some very large radii.

The physical parameterizations include WRF Single-moment 6-class scheme (Hong and Lim 2006) for microphysics processes, the rapid radiative transfer model for general circulation models (RRTMG) scheme (Iacono et al. 2008) for both short-wave and long-wave radiations, Yonsei University (YSU) Scheme (Hong et al., 2006) for planetary boundary layer processes, and the fifth-generation Pennsylvania State University/National Center for Atmospheric Research Mesoscale Model (MM5) similarity scheme (Zhang and Anthes 1982) for surface layer physics. Considering that the 4-km-resolution inner domain is large enough to cover most structure of the simulated TCs and the two-way interactive nesting is applied, no cumulus scheme is employed in either domain.

* 1. **Experiment design**

Two groups of experiments are designed. We analyze the responses of TC intensity, outer size and destructive potential to dry midlevel air in the first group. The second group will be used to test the sensitivity of the assumptions in the first group. We will focus on the first group further to investigate the physical mechanism of the outer size growth as TCs in sensitivity experiments in both groups show qualitatively consistent changes to the dry midlevel air insertion. The detailed description of these two groups of experiments is given as follows.

In the first group, the CTRL run is set up by following the model specification in Section 2.1. As for the sensitivity experiments, a dry midlevel air layer is firstly inserted only outside the 400-km radius from the cyclone center at both developing stage (experiment D\_400KM) and mature stage (experiment M\_400KM). The moist envelope is then removed so that the dry air influences the cyclone inner core directly at the developing stage (experiment D\_ALL) and mature stage (experiment M\_ALL).

In the second group, we repeat CTRL and four sensitivity runs in the first group but (1) with the Mellor-Yamada-Janjic scheme for planetary boundary layer processes, (2) with the Lin et al*.* (1983) scheme for microphysics processes, (3) with 31 σ-levels, (4) using triply nested domains with 1.33-km grid spacing in the finest nesting, (5) perturbing the midlevel temperature so that the horizontal gradient of virtual temperature created by the RH perturbation can be counteracted, and (6) using an initial sounding profile that has been adjusted to the radiative-convective equilibrium (RCE) state. The equilibrated sounding is adopted from (Wang and Toumi 2018a), which is based on the Jordan hurricane season mean sounding but adjusted towards the RCE state on an *f*-plane (20oN) with an SST of 27oC and an extra surface zonal wind set to 5 m s-1. The second group consists of 30 experiments in total, 24 of which are sensitivity runs.

After the initial spinup, the dry midlevel air is inserted at simulation hour 72 and 144, respectively, and then the simulation is restarted. As will be shown in Section 3, simulation hour 72 is during the developing stage while simulation hour 144 is during the mature stage. The dry midlevel air is inserted at these two times so that we can investigate whether the dry midlevel air can influence the TC outer circulation in different ways when the TC structure changes from development to maturity. We set the radius of the moist envelope to 400 km as Braun (2010) showed that the dry midlevel air is located typically more than 400 km away from the storm center.

A single consistent nomenclature is used for referencing time, e.g., “T72” for simulation hour 72 in a non-restart run, and “T72+24” in a restart run for simulation hour 24 after a 72-hour restart. All the restart runs last for 3 days so that the mean TC changes in all the sensitivity runs can be compared. Kimball (2006) showed that real TCs can evolve very differently during the next 36 hours after the environmental humidity is perturbed. Three days in our restart runs should, therefore, be long enough to simulate the TC response to the presence of dry midlevel air.

We perturb the midlevel air RH to 40% by changing the mixing ratio only and will show results for this change. We have also changed the midlevel RH to 20% and 60% as sensitivity tests. The midlevel air is defined between 850-600 hPa (Braun et al. 2012). Figure 1a shows the Jordan hurricane season mean sounding, the RH of which between 850-600 hPa (1547-4442 m in height) varies from 74% to 50%. As shown in Figure 1a, Jordan (1958) did not report any humidity data above 450 hPa (6703 m in height) due to the limitation of radiosondes on detecting some very low humidity at that time. In the model setup, the missing specific humidity data is initially set to 0 g kg-1. It should be noted that the midlevel air in the tropics is featured by low θe air by nature (Figure 1a). To some extent, inserting dry midlevel air mass at a certain distance from the TC center is physically plausible. We note that an RH of about 40% in the mid-level is also consistent with observations of the Saharan air layer (Braun 2010) and we will discuss its potential impact on the outer size of Atlantic hurricanes.

Figure 1b and 1d show the vertical profiles of RH, temperature, and equivalent potential temperature (θe) at T72 and T144 in CTRL of the first group, which are averaged within 400-800 km radius. The calculation of θe follows Bolton (1980) for a water-saturation pseudo-adiabatic process. Figure 1b and 1d show that from T72 to T144 the temperature sounding is almost identical. The vertical RH profile at these times are similar, but a slight moistening with time can be seen in the upper troposphere. The grey lines in Figure 1 show the 9th to the 14th σ-levels that are located from 577 hPa to 832 hPa in the model. The RH at these σ-levels is perturbed in the sensitivity runs. As shown in Figure 1b, the midlevel RH at T72 in CTRL is about 60-75%. Figure 1c and 1e show the modified RH and θ­­e soundings at T72+0 in D\_400KM and at T144+0 in M\_400KM, respectively, after the mixing ratio perturbation. The temperature profiles in Figure 1c and 1e are not changed.

Our sensitivity experiments are reminiscent of Hill and Lackmann (2009), but the differences from their work are that (1) the RH is only modified in the midtroposphere instead of the whole air column, and (2) the dry air is further away from the core convection [400 km in radius rather than 100 km as in Hill and Lackmann (2009)]. Extra experiments analogous to Hill and Lackmann (2009) are also conducted to give the reader more confidence in the model setup in this study. Based on the experiment D\_400KM and M\_400KM, we extend the dryness to the whole air column and move the edge of the inserted dry air inwards from 400 km to 100 km in radius to the cyclone center, which is similar to the configuration used by Hill and Lackmann (2009).

The passive tracer option in WRF is activated to study the potential pathways of midlevel air. The time and locus of tracer releasing will be introduced in the next section. The passive tracer has no impact on simulations. The tracers are advected and diffused by the local dynamics.

**3. Results**

We first analyze the results from the first group of experiments. The intensity is measured by the maximum wind speed (Vm) based on the azimuthally averaged wind profile at 10 m. Figure 2 shows the life cycle of Vm in CTRL and four sensitivity runs. The Vm in CTRL increases rapidly in the first two simulation days. The intensity in the third day is equivalent to a Category-1 cyclone. There is another rapid intensification in CTRL between T96 and T120. The Vm in CTRL reaches a relatively steady stage after T144. Based on the intensity evolution, we divide the whole life cycle into three periods: (1) from T0 to T72, (2) from T72 to T144, and (3) from T144 to T216. The first period is taken as the initial spinup. The dry midlevel air is placed in the second and third periods, respectively, to study its impact on the TC outer circulation during the developing and mature stages.

After inserting the dry midlevel air, Vm drops steeply between T72+0 and T72+12 in D\_ALL and between T144+0 and T144+12 in M\_ALL (Figure 2), in which the moist envelope does not protect the TC core. When a moist envelope exists during the developing stage (e.g., D\_400KM in Figure 2), it takes about 36 hours (around T108) for the environmental dry midlevel air to start to slow down the intensification process. However, at the end of the three-day restart runs, D\_ALL and D\_400KM can still reach a similar lifetime maximum intensity as in CTRL.

The Hovmöller diagram of the azimuthally averaged 10-m wind speed in CTRL shows a continuous expansion of the outer circulation (Figure 3a, 3b and 3e). In the sensitivity experiments, we see small but immediate decreases of 10-m wind speed in the outer circulation after placing the dry air throughout the entire midtroposphere (D\_ALL in Figure 3d and M\_ALL in Figure 3g). On the other hand, if the TC core is protected by a moist envelope (D\_400KM in Figure 3c and M\_400KM in Figure 3f), it is surprising that the environmental dry midlevel air leads to a considerable broadening of the outer circulation. It takes about a day for the dry midlevel air to have an impact on R18 in D\_400KM, and R18 is 75 km or 55% larger than CTRL after two days of inserting the dry midlevel air.

Other interesting results can also be seen in Figure 3. Firstly, when the cyclone center is protected by a moist envelope, the increase of 10-m wind speed mainly happens within a radius of 400 km (the vertical dashed lines in Figure 3c and 3f), which is the initial edge of the inserted dry midlevel air. Secondly, the positive wind anomalies in Figure 3c and 3f show an inward propagation towards the radius of maximum wind speed (RMW). The impacts of environmental dry midlevel air on TC outer circulation are more pronounced during the developing stage (Figure 3c) than the mature stage (Figure 3f). The wind anomaly outside the radius of 100 km in M\_400KM returns to zero around T144+60 (Figure 3f). This is mainly due to a more rapid expansion of the outer wind field in CTRL for T144-T216 than for T72-T144. Moreover, although the dry midlevel air influences the TC inner core in Figure 3d, the outer circulation starts to recover at T72+48 when some positive wind anomalies appear around the radius of 100-200 km.

Our sensitivity runs of changing RH to 20% and 60% show the same qualitative behavior. The response appears somewhat non-linear with enhanced sensitivity to drying. Here we are concerned with basic processes rather than the quantitative response of these idealized simulations. In a further run we placed the column dry air outside a radius of 100 km during the early and late stages. With this configuration, we can qualitatively reproduce the results in Hill and Lackmann (2009) during both developing and mature stages, i.e., the TC outer circulation shrinks if the column dry insertion is close enough to the core.

We next analyze the changes in TC destructive potential by calculating the Integrated Power Dissipation (IPD) index (Emanuel 2005). The calculation of IPD follows Wang and Toumi (2016), which is rewritten here as,

 (2)

where *ρ* is the air density set as 1.1 kg m-3, *CD* is the drag coefficient calculated with wind speed (Large and Yeager 2008), *V* is the azimuthally averaged wind speed at 10 m, and *S* is the integral area with a wind speed of at least gale force. As a combined effect of the changes in intensity and outer circulation, the TC destructive potential (IPD) decreases if the dry midlevel air reaches the eyewall convection directly (D\_ALL and M\_ALL in Figure 4), whereas IPD increases if the dry midlevel air is located away from the TC center (D\_400KM and M\_400KM in Figure 4).

Figure 5 displays the responses of intensity (Vm), outer size (R18) and destructive potential (IPD) in both groups of experiments. For ease of comparison, the experiments in the first group and the mean of two groups are highlighted with thick lines. Figure 5 shows a common evolution in both groups, for example, the expected immediate reduction of intensity if the dry midlevel air invades the convective core (Figure 5a), and, importantly for this study, the increases of outer size and consequently the destructive potential if the dry midlevel air is only inserted at a large radius initially (Figure 5b and in 5c). During the developing stage, the mean destructive potential can be more than doubled after three days of dry insertion (Figure 5c). This increase in destructive potential is mainly due to the broadening of the outer circulation as shown in Figure 5b. The recovery of the outer size can also be seen in the mean simulation (blue lines in Figure 5b) as found in Figure 3d.

The TCs in both groups show qualitatively consistent responses to dry midlevel air during the developing and mature phases. The response is robust for the extensive range of sensitivities of model resolution and physics packages examined. We next only focus on CTRL, D\_400KM and D\_ALL in the first group to analyze the physical linkage between the dry midlevel air and the outer circulation change.

Figure 6 shows θe and RH in CTRL, D\_400KM and D\_ALL, respectively, at T72+0, T72+24, T72+48, and T72+72. The environmental midlevel air in CTRL is already characterized by low θe ­(Figure 6a, 6d, 6g, and 6j). After inserting the dry midlevel air, the vertical contrast of θe is further increased (Figure 6b and 6c). The differences of both θe and RH distributions among these three runs decrease gradually with time.

It is worth noting that at T72+24 the midlevel RH in D\_400KM increases between the radius of 100-250 km (Figure 6e). The midlevel high-RH structure becomes separated from the humid eyewall after T72+48 according to the evolution of the 85% RH contour (Figure 6h and 6k). This structure with high RH weakens from T72+24 (Figure 6e) to T72+72 (Figure 6k). A similar but much weaker increase in the midlevel RH also happens in D\_ALL around a radius of 100 km but at a later stage (T72+48, Figure 6i). Interestingly, a positive 10-m wind anomaly in D\_ALL also firstly appears at a radius of 100 km in Figure 3d atthe same time.

Figure 6 shows that the vertical RH distribution in the rainbands can be altered by dry midlevel air intrusion from the ambient environment. Next, we apply passive tracer analyses to track the movement of the initial dry midlevel air in detail. In CTRL, D\_400KM and D\_ALL, tracers are released at a radius of 400 km from the TC center at T72 when the dry midlevel air is inserted. The initial tracer concentration is set to unity. Tracers are initially released at about 780 hPa. In D\_400KM, tracers are located on the edge of the inserted dry air. Two different tracer configurations are applied: the “ring tracers” and the “point tracers”. The former ones are a ring of tracers surrounding the TC center, and the latter are four point tracers located to the north, south, west and east of the cyclone center.

Figure 7 shows the concentration of point tracers at T72+24 in D\_400KM at about 200 hPa, 500 hPa, 780 hPa and near the surface. Four markers in Figure 7c show the initial loci of four point tracers. After releasing the tracers on the edge of inserted dry midlevel air, the tracers wrap cyclonically inwards (Figure 7c). The tracers with similar motions are also found at 500 hPa (Figure 7b) and near the surface (Figure 7d). The tracers in the boundary layer (Figure 7d) appear to move faster towards the eyewall than in the midtroposphere (Figure 7c). Twenty-four hours after the tracer releasing, a considerable number of tracers are found at 200 hPa (Figure 7a). The movements of four point tracers are similar, which enables us to analyze the dry midlevel air pathways with the azimuthally accumulated concentration of ring tracers.

Figure 8 displays the azimuthally accumulated concentration of ring tracers in CTRL, D\_400KM and D\_ALL. The tracers in D\_400KM have largely mixed in the vertical at T72+24 with relatively slow but robust inward movement (Figure 8e). A large portion of the tracers has reached the inflow layer. It is worth noting that atthe same time, a certain number of tracers also appear in the outflow layer even before the high concentration mass reaches the eyewall region through the boundary layer. In the next 24 hours (T72+48, Figure 8h), the tracers at middle and lower levels continue to move inwards. A vertical tracer conveyor belt connecting the inflow and outflow layers, as shown in Figure 8e, can also be identified between the radius of 100-300 km in Figure 8h. After 72 hours of tracer releasing (T72+72, Figure 8k), most tracers are accumulated in the outflow layer at large radius. Those tracers would then descend slowly within weeks due to the effect of radiative cooling.

Similar tracer pathways can be found among CTRL, D\_400KM and D\_ALL in Figure 8. For example, the vertical conveyor belt in the rainbands exists in CTRL (Figure 8d, 8g and 8j) and D\_ALL (Figure 8i and 8l) as well. However, the vertical spread of tracers is faster in D\_400KM than the other two in the rainbands, which indicates that the dry midlevel air intrusion changes the outer radial-vertical overturning circulation. The efficient vertical spread of tracers away from the eyewall in D\_400KM is attributed to the convective rainbands as shown in Figure 9. We next analyze in detail to explore the physical mechanism for the outer circulation expansion due to the dry air intrusion from the midtroposphere.

Figure 10 shows the latent heating rate from the microphysics scheme in CTRL, D\_400KM and D\_ALL at T72+6, T72+24, and T72+72. The microphysics latent heating is calculated as the net effect of condensation heating and evaporative cooling. Six hours after the dry air insertion (T72+6), the latent heating in D\_400KM (Figure 10b) is largely suppressed outside the 400-km radius where the midlevel RH is perturbed to 40%. Importantly, the latent heating within the edge of the inserted dry air appears to be enhanced. However, in D\_ALL (Figure 10c), the condensation heating and evaporative cooling are both reduced across the domain. We can see a marked horizontal gradient of latent heating around the radius of 400 km in D\_400KM, but not D\_ALL. Twenty-four hours after the dry air insertion (T72+24), the horizontal gradient of latent heating in D\_400KM is further increased (Figure 10e). At the same time, the latent heating outside 400-km radius in both D\_400KM and D\_ALL starts to be reinvigorated above the boundary layer (Figure 10e and 10f). At T72+72, the differences of latent heating in CTRL (Figure 10g), D\_400KM (Figure 10h) and D\_ALL (Figure 10i) are much reduced.

Figure 11a shows an evident gradient of the latent heating anomaly around the 400-km radius which is the edge of the dry intrusion in D\_400KM. A least squares linear fitting of the radial heating anomaly at the height of 4 km gives a heating gradient of about -4 K day-1 per 100 km from the radius of 300 to 500 km.

Removing substantial amounts of water vapor has also radiative heating effects. In the mid layer where the dry air is inserted, the combined effect of short and long-wave radiative effects is a small net radiative cooling of only about -0.3 K day-1 per 100 km at the height of 4 km from 300 km to 500 km in radius, and the gradient of diabatic heating generated by latent heating (Figure 11a) around the radius of 400 km is, therefore, one order of magnitude larger. In other words, the latent heating is the dominant factor on the anomalous heating gradient.

According to the Sawyer-Eliassen equation, a horizontal gradient of heating (for example, the outward decrease in latent heating) can create an overturning circulation (e.g., Willoughby 1979). Indeed, Figure 11c shows the transverse motions of a forced overturning gyre anomaly in D\_400KM near the radius of 400 km where the heating gradient is large. The vertical velocity anomaly at a height of 5 km and averaged between 200 to 400 km from the TC center is +0.02 m s-1in Figure 11e. It is important to note that the anomalous overturning circulation in Figure 11c overlaps with the boundary layer, which accelerates the inflow speed of the main secondary circulation. A stronger boundary layer inflow favors the angular momentum accumulation in the rainbands, which makes the TC outer circulation grow (Chan and Chan 2013).

In addition, the radial inflow anomaly in Figure 11c also extends into the eyewall region, and correspondingly, we can see another increase of vertical velocity within the radius of 100 km in Figure 11e. However, the largest increment in Figure 11e is still an order of magnitude less than the vertical velocity in CTRL (1.2 m s-1), so the intensity in D\_400KM and CTRL are almost identical in the first six hours after the dry intrusion. Further analysis (not shown) reveals that the strong outflow anomaly in Figure 11c happens at the height where the main outflow is. This means both the outgoing and ingoing legs of the anomalous overturning circulation are connected with the main secondary circulation. This circulation also promotes convection and heating that further amplifies the horizontal heating gradient.

On the contrary, this gradient of heating anomaly does not exist in D\_ALL (Figure 11b) during the same period as shown in Figure 11a. Figure 11b also shows a considerable reduction of latent heating in the vicinity of the eyewall. This corresponds to a slowdown of the secondary circulation as found in Figure 11d and 11f, which is due to the dry air intrusion directly into the inner core in D\_ALL. The weakening of vertical velocity in D\_ALL (Figure 11f) is comparable to the magnitude of the eyewall ascending in CTRL, and thus there is a rapid reduction of intensity in D\_ALL after the dry air insertion.

Our tracer analyses have shown that the initial dry midlevel air can be vertically entrained into the boundary layer and then flushes inwards (e.g., Figure 7 and 8). The mixed dry air can increase the near-surface thermodynamic disequilibrium and then may change the surface enthalpy flux. The thermodynamic disequilibrium here is quantified as the difference between the enthalpy of the ocean surface (ks) and the enthalpy at the lowest σ-level (ka). Figure 12a shows an enhanced ks-ka of at least 1600 J in D\_400KM at about 200-350 km in radius before T72+12. There is a clear inward change of positive ks-ka anomaly from a radius of 350 to 100 km, which is denoted by the black arrow in Figure 12a. Note that this inward change only happens from T72+12 to T72+36. The strength of this enthalpy contrast increases with the inward change. Similarly, the enthalpy flux in D\_400KM (Figure 12c) also shows an inward increase along the black arrow. Due to the increase in the surface flux, the mixed dry air from the midtroposphere is “recharged” with high enthalpy. This recharging process is essential to the maintenance of the eyewall convection. In addition, there is a weak outward propagation of positive ks-ka in T72+36-T72+48 from a radius of 100 to 400 km, which appears to follow the expansion of the wind field during that period.

In D\_ALL (Figure 12b), the thermodynamic disequilibrium is even larger than D\_400KM, but the increase in the surface enthalpy flux is much less in Figure 12d. This is because of the decrease in surface wind speed as shown in previous analyses. Nevertheless, the enthalpy contrast is so strong that the surface enthalpy flux anomaly in D\_ALL (Figure 12d) is still positive between RMW and R18 after about T72+30. This continuous input of enthalpy from the warm ocean is one of the reasons that the whole circulation in D\_ALL can recover to some extent after three days of dry intrusion.

The physical processes, related to the environmental dry midlevel air ventilation and the consequent expansion of the TC outer primary circulation, are illustrated schematically in Figure 13. The dry midlevel air can be ventilated progressively into the outer rainbands. The direct effect of dry air intrusion in the midtroposphere is the depression of local moist convection, and it leads to a local decrease of latent heating. This enhances the horizontal gradient of heating in the rainbands, which initiates an anomalous radial-vertical overturning circulation. The ascending leg of the outer overturning circulation enhances the condensation heating locally, and this further increases the horizontal gradient of heating and thus strengthens the outer overturning circulation. Based on our simulation, the absolute heating anomalies on the inward side of the edge of the insert dry air are stronger and more concentrated than those on the outward side. We indicate these characteristics in Figure 13 by using a small area with a large “+” sign, and a large area with a small “-” sign. The outgoing leg of the outer overturning circulation merges with the outflow of the main secondary circulation. The ingoing leg overlaps with the inflow layer of the main secondary circulation. One part of the ingoing leg turns into the ascending leg, and the other part joins the major inflow and flushes towards the eyewall. The outer overturning circulation increases the angular momentum import by accelerating the outer radial inflow in the boundary layer, and thus the TC outer size increases.

The second process can be regarded as a “recharge” mechanism. Any midlevel dry air mixed into the boundary layer increases the near-surface thermodynamic disequilibrium. Without any strong decrease in the low-level wind speed, a large vertical enthalpy contrast enhances the surface enthalpy flux. The entrained dry air into the inflow layer is “recharged” with high enthalpy before approaching the eyewall. The recharge process sustains the enthalpy input to the TC system. Even if the dry air invades the inner core directly, in our simulation, the circulation can still recover through the recharging process.

**4. Discussion**

The influence of dry midlevel air on TC core intensity in our simulations is consistent with previous studies. For example, we show that the dry midlevel air can slow down the intensification rate during the developing stage in idealized simulations, which agrees with the finding with real TCs in Kaplan and DeMaria (2003). Our results also show that the TC influenced by dry midlevel air can still reach a similar intensity as CTRL but it takes a longer time, which is in line with Braun et al. (2012).

Our experiments show that the outer circulation of an existing TC can be expanded considerably if the midlevel dry air is placed 400 km away from the center. The mean of sensitivity experiments also shows that the outer size, for example, measured by R18, can also be more than 10% larger than CTRL after three days of the dry air insertion directly into the core (the thick blue line at T144 inFigure 5b). Braun (2010) showed that the Saharan dry midlevel air is located typically more than 400 km away from the storm center. Our results suggest that even if the dry midlevel air is transported closer to the center than 400 km, it is expected to have a similar expansion of the outer circulation.

Hill and Lackmann (2009) found that the whole wind field expands in an initially moist environment due to a broadening of the potential vorticity tower by diabatic heating. Their conclusion seems to be contradictory to our results, but the physical mechanism behind the broadening of the wind field in these two studies agrees with each other. Both studies emphasize the important role of latent heating in the rainbands in expanding the outer circulation.

The vital ingredient of our mechanism to explain the outer size growth is the horizontal gradient of latent heating in the rainbands. According to the Sawyer-Eliassen equation, the horizontal gradient of the heating and the vertical derivative of the momentum convergence are two terms that can potentially force an azimuthal-mean secondary circulation (Willoughby 1979). The anomalous radial-vertical overturning circulation in our simulation is a response to the enhanced horizontal gradient of latent heating due to the dry insertion. As mentioned in Section 2, we repeat the first group of experiments but adjusting the midlevel temperature to counteract the virtual temperature gradient created by the RH perturbation. The sensitivity set with adjusted virtual temperature shows very similar changes in Vm, R18 and IPD as found in the first group. These sensitivity experiments confirm the heating gradient as the main driver of the outer solenoidal circulation, rather than the temperature gradient. Furthermore, the lower inertial stability in the outer weaker wind regions (compared to the core) provides less resistance to the radially inward motion in response to the heating gradient. In other words, the radial heating gradient change can be relatively small at large radii (compared to near the core), but there will be a response of radial momentum inflow in this low inertial stability region (Chan et al. 2019).

The proposed mechanism may shed some light on the continuous expansion of the outer circulation during the mature stage. It has long been recognized that the TC outer circulation expands after its maturity (Riehl 1954). Wang and Toumi (2017) found that about a quarter of global TCs showed a continuous broadening of the outer circulation for at least one day after reaching the lifetime maximum intensity. As expected, the TC wind profile in this study also expands throughout the whole lifecycle (Figure 3a, 3b, and 3e). The extra dry midlevel air in our sensitivity experiments is added to the vertical layer where θe is already the minimum in the vertical (Figure 1). The evolution of tracer concentration is similar in CTRL and D\_400KM as shown in Figure 8. The main difference is that the tracer in D\_400KM, where the midlevel air at large radius dries more, shows the more rapid spread than that in CTRL. All these results suggest that the proposed physical mechanism linking the dry midlevel air and the expansion of the TC outer circulation could be a common process. This process is just amplified in our experimental design.

The proposal physical mechanism may also be pertinent to the potential influence of the Saharan dry air layer on the lifetime maximum size of Atlantic hurricanes. If the Saharan air layer is located away from the TC center at an early stage (e.g., 400 km away), it can increase the horizontal heating gradient in the rainbands and expand the outer circulation. Assuming that this anomalous size growth is maintained in real hurricanes (as shown in Figure 5b), it would provide an extra boost to the outer size growth, which is not available in other basins, and may thus contribute to creating a larger lifetime maximum size of cyclones in the Atlantic. We will examine this hypothesis with observations and simulations in a further study.

It is also interesting to note that the ascending leg of the forced overturning circulation is initially separated from the eyewall (Figure 11e). This ascending leg contracts and eventually merges with the eyewall updraft (Figure 3c and 3f). A moistened region corresponding to the ascending leg of the outer overturning circulation is found in Figure 6e and 6h. A delayed intensification in D\_400KM is also shown in Figure 2. All these characteristics are reminiscent of eyewall replacement cycles. However, there are two marked differences from the eyewall replacement. Firstly, in our simulation, the original eyewall survives during the merging process, rather than dies (Houze et al. 2007). Secondly, we do not see any double peaks in the low-level wind profiles near RMW.

Two pathways for the dry midlevel air moving towards the eyewall are identified in this study. They are partly consistent with the concept of midlevel ventilation pathways by Tang and Emanuel (2010) and Riemer et al*.* (2010). Tang and Emanuel (2010, 2012) illustrated that the first pathway is through the downdraft outside the eyewall and the sub-cloud layer. The second pathway is horizontal. In our simulations, the downwards spread of tracers in the first pathway may not be caused by the eyewall downdraft as it happens at least 300 km away from the TC center. According to the tracer analyses, the dry air is mixed rapidly in the vertical. Besides, our simulations show that a large number of tracers reach the outflow layer by local convection in the rainbands, rather than passing through the eyewall as described in the previous midlevel ventilation framework (Tang and Emanuel 2010, 2012). It is important to note that the vertical wind shear, which is important to dry air ventilation into TCs, is absent in our simulations. Thus, more detailed work may be required to generalize the findings of this study in a vertically sheared environment.

Recent studies also showed that the cyclone can be influenced by its own midlevel dry air via similar pathways as found in our simulations. For example, Alland et al. (2017) showed that the low-θe air from the midtroposphere can subside into the boundary layer between the radius of 200 and 400 km in an axisymmetric TC model. Another example is given by Corsaro and Toumi (2017) who found that the idealized TC with β-drift can recirculate low-entropy air back into the tropical cyclone from upper and middle levels.

The difference between our work and other previous studies is that we insert the dry midlevel air into the environment after the initial spinup, rather than placing the dry midlevel air before the TC genesis (Braun et al., 2012) or drying the whole air column (Kimball 2006; Hill and Lackmann 2009). Certain conditions are therefore required for our physical mechanism of dry midlevel air promoting the expansion of the outer circulation: (1) a TC that is spun up; (2) a warm ocean under the whole or most part of the TC; (3) dry midlevel air away from the core convection. In addition, it is unlikely for a developing or mature TC to be encircled completely by dry midlevel air as has been tested in this idealized study. This process, therefore, needs to be further examined with real TC cases, or with idealized simulations but injecting the dry midlevel air in only part of the outer circulation.

**5. Conclusions**

In this study, we aim to answer the question: can the dry midlevel air from the ambient environment change the outer primary circulation? We find that the environmental dry midlevel air can broaden the outer primary circulation in both developing and mature stages. This finding is qualitatively robust based on 28 sensitivity experiments. When the relative humidity of dry midlevel air outside a radius of 400 km from the cyclone center is reduced from a typical value to 40% at the developing stage, our simulations show that the mean integrated power dissipation more than doubles after three days.

Further analyses show that the expansion of the TC outer circulation is due to the enhanced horizontal gradient of latent heating in the rainbands. The dry air intrusion in the midtroposphere leads to a local decrease of latent heating. This enhances the horizontal gradient of heating in the rainbands, which initiates an anomalous radial-vertical overturning circulation at large radius. The ascending leg of the outer overturning circulation enhances the condensation heating locally, and this further increases the gradient of heating and thus strengthens the outer overturning circulation. The ingoing leg of the outer overturning circulation overlaps with the inflow layer of the main secondary circulation. This increases the angular momentum import by accelerating the outer radial inflow in the boundary layer. Therefore, the TC outer size increases.

The other important part of the proposed mechanism can be regarded as a “recharge” process. Any midlevel dry air mixed into the boundary layer increases the near-surface thermodynamic disequilibrium, which enhances the surface enthalpy flux. The entrained dry air into the inflow layer is replenished or “recharged” with high enthalpy when flushing towards the eyewall. This process sustains the enthalpy input. According to our simulations, the TC circulation can still recover through the recharging process even if the dry air invades the inner core directly.

Apart from any potential applications to the TC wind structure forecasting, the proposed physical mechanism between the environmental dry midlevel air and cyclone growth may explain the continuous expansion in TC outer circulation after maturity as found in observations.

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**Reference**

Alland, J. J., B. H. Tang, and K. L. Corbosiero, 2017: Effects of Midlevel Dry Air on Development of the Axisymmetric Tropical Cyclone Secondary Circulation. *J. Atmos. Sci.*, **74**, 1455–1470, doi:10.1175/JAS-D-16-0271.1. http://journals.ametsoc.org/doi/10.1175/JAS-D-16-0271.1.

Bolton, D., 1980: The Computation of Equivalent Potential Temperature. *Mon. Weather Rev.*, **108**, 1046–1053, doi:10.1175/1520-0493(1980)108<1046:TCOEPT>2.0.CO;2. http://journals.ametsoc.org/doi/abs/10.1175/1520-0493%281980%29108%3C1046%3ATCOEPT%3E2.0.CO%3B2 (Accessed May 10, 2018).

Braun, S. A., 2010: Reevaluating the Role of the Saharan Air Layer in Atlantic Tropical Cyclogenesis and Evolution. *Mon. Weather Rev.*, **138**, 2007–2037, doi:10.1175/2009MWR3135.1. http://journals.ametsoc.org/doi/abs/10.1175/2009MWR3135.1.

——, J. A. Sippel, and D. S. Nolan, 2012: The Impact of Dry Midlevel Air on Hurricane Intensity in Idealized Simulations with No Mean Flow. *J. Atmos. Sci.*, **69**, 236–257, doi:10.1175/JAS-D-10-05007.1. http://journals.ametsoc.org/doi/abs/10.1175/JAS-D-10-05007.1.

Chan, K. T. F., and J. C. L. Chan, 2013: Angular Momentum Transports and Synoptic Flow Patterns Associated with Tropical Cyclone Size Change. *Mon. Weather Rev.*, **141**, 3985–4007.

——, D. Wang, Y. Zhang, W. Wanawong, M. He, and X. Yu, 2019: Does strong vertical wind shear certainly lead to the weakening of a tropical cyclone? doi:10.1088/2515-7620/aaecac. https://doi.org/10.1088/2515-7620/aaecac (Accessed March 29, 2019).

Corsaro, C. M., and R. Toumi, 2017: A self-weakening mechanism for tropical cyclones. *Q. J. R. Meteorol. Soc.*, **143**, 2585–2599, doi:10.1002/qj.3109. http://doi.wiley.com/10.1002/qj.3109.

DeMaria, M., J. a. Knaff, and B. H. Connell, 2001: A Tropical Cyclone Genesis Parameter for the Tropical Atlantic. *Weather Forecast.*, **16**, 219–233, doi:10.1175/1520-0434(2001)016<0219:ATCGPF>2.0.CO;2.

Ditchek, S. D., T. C. Nelson, M. Rosenmayer, and K. L. Corbosiero, 2017: The Relationship between Tropical Cyclones at Genesis and Their Maximum Attained Intensity. *J. Clim.*, **30**, 4897–4913, doi:10.1175/JCLI-D-16-0554.1. http://journals.ametsoc.org/doi/10.1175/JCLI-D-16-0554.1.

Emanuel, K., 2005: Increasing Destructiveness of Tropical Cyclones over the Past 30 Years. *Nature*, **436**, 686–688.

Emanuel, K. A., and D. S. Nolan, 2004: Tropical cyclone activity and the global climate system. *Preprints, 26th Conf. on Hurricanes and Tropical Meteorology, Miami, FL, Amer. Meteor. Soc. A*, Vol. 10 of.

Gray, W. M., 1968: Global View of the Origin of Tropical Disturbances and Storms. *Mon. Weather Rev.*, **96**, 669–700, doi:10.1175/1520-0493(1968)096<0669:GVOTOO>2.0.CO;2. http://journals.ametsoc.org/doi/abs/10.1175/1520-0493%281968%29096%3C0669%3AGVOTOO%3E2.0.CO%3B2.

Gray, W. M., 1975: Tropical Cyclone Genesis. *Dept. Atmos. Sci. Colarado State Univ. Fort Collins, CO, 121.*, **Paper No.**, 120.

Hill, K. A., and G. M. Lackmann, 2009: Influence of Environmental Humidity on Tropical Cyclone Size. *Mon. Weather Rev.*, **137**, 3294–3315.

Hong, S.-Y., and J.-O. J. Lim, 2006: The WRF Single-Moment 6-Class Microphysics Scheme (WSM6). *J. Korean Meteorol. Soc.*, **42**, 129–151.

——, Y. Noh, and J. Dudhia, 2006: A New Vertical Diffusion Package with an Explicit Treatment of Entrainment Processes. *Mon. Weather Rev.*, **134**, 2318–2341.

Houze, R. a, S. S. Chen, B. F. Smull, W.-C. Lee, and M. M. Bell, 2007: Hurricane Intensity and Eyewall Replacement. *Science (80-. ).*, **315**, 1235–1239, doi:10.1126/science.1135650. http://www.sciencemag.org/cgi/doi/10.1126/science.1135650.

Iacono, M. J., J. S. Delamere, E. J. Mlawer, M. W. Shephard, S. A. Clough, and W. D. Collins, 2008: Radiative forcing by long-lived greenhouse gases: Calculations with the AER radiative transfer models. *J. Geophys. Res. Atmos.*, **113**, 2–9, doi:10.1029/2008JD009944.

Jordan, C., 1958: Mean Soundings for the West Indies Area. *J. Meteorol.*, **15**, 91–97.

Kaplan, J., and M. DeMaria, 2003: Large-Scale Characteristics of Rapidly Intensifying Tropical Cyclones in the North Atlantic Basin. *Weather Forecast.*, **18**, 1093–1108, doi:10.1175/1520-0434(2003)018<1093:LCORIT>2.0.CO;2.

Kimball, S. K., 2006: A modeling study of hurricane landfall in a dry environment. *Mon. Weather Rev.*, **134**, 1901–1918, doi:10.1175/MWR3155.1.

Knapp, K. R., M. C. Kruk, D. H. Levinson, H. J. Diamond, and C. J. Neumann, 2010: The International Best Track Archive for Climate Stewardship (IBTrACS). *Bull. Am. Meteorol. Soc.*, **91**, 363–376, doi:10.1175/2009bams2755.1.

Large, W. G., and S. G. Yeager, 2008: The Global Climatology of an Interannually Varying Air–Sea Flux Data Set. *Clim. Dyn.*, **33**, 341–364.

Lin, Y.-L., R. D. Farley, and H. D. Orville, 1983: Bulk Parameterization of the Snow Field in a Cloud Model. *J. Clim. Appl. Meteorol.*, **22**, 1065–1092, doi:10.1175/1520-0450(1983)022<1065:BPOTSF>2.0.CO;2. http://journals.ametsoc.org/doi/abs/10.1175/1520-0450%281983%29022%3C1065%3ABPOTSF%3E2.0.CO%3B2.

Nolan, D. S., 2007: What is the trigger for tropical cyclogenesis? *Aust Meteorol Mag*, **56**, 241–266.

Riehl, H., 1954: *Tropical meteorology*. McGraw-Hill,.

Riemer, M., M. T. Montgomery, and M. E. Nicholls, 2010: A new paradigm for intensity modification of tropical cyclones: thermodynamic impact of vertical wind shear on the inflow layer. *Atmos. Chem. Phys.*, **10**, 3163–3188, doi:10.5194/acp-10-3163-2010.

Simpson, R., and R. Riehl, 1958: Mid-tropospheric ventilation as a constraint on hurricane development and maintenance. *Preprints, Tech. Conf. on Hurricanes, Miami Beach, FL, Amer. Meteor. Soc*, D4-1-D4-10.

Skamarock, W. C., and Coauthors, 2008: A Description of the Advanced Research WRF Version 3. *Tech. Rep.*, 113, doi:10.5065/D6DZ069T.

Tang, B., and K. Emanuel, 2010: Midlevel Ventilation’s Constraint on Tropical Cyclone Intensity. *J. Atmos. Sci.*, **67**, 1817–1830, doi:10.1175/2010JAS3318.1.

——, and ——, 2012: A ventilation index for tropical cyclones. *Bull. Am. Meteorol. Soc.*, **93**, 1901–1912, doi:10.1175/BAMS-D-11-00165.1.

Wang, S., and R. Toumi, 2016: On the relationship between hurricane cost and the integrated wind profile. *Environ. Res. Lett.*, **11**, 114005, doi:10.1088/1748-9326/11/11/114005. http://stacks.iop.org/1748-9326/11/i=11/a=114005?key=crossref.b8c4fd8906504a99c87f1218d5c8042f.

——, and ——, 2018a: Reduced Sensitivity of Tropical Cyclone Intensity and Size to Sea Surface Temperature in a Radiative-Convective Equilibrium Environment. *Adv. Atmos. Sci.*, **35**, 981–993, doi:10.1007/s00376-018-7277-5. https://link.springer.com/content/pdf/10.1007%2Fs00376-018-7277-5.pdf (Accessed June 7, 2018).

——, and ——, 2018b: A historical analysis of the mature stage of tropical cyclones. *Int. J. Climatol.*, **38**, 2490–2505, doi:10.1002/joc.5374. http://doi.wiley.com/10.1002/joc.5374.

——, ——, A. Czaja, and A. Van Kan, 2015: An analytic model of tropical cyclone wind profiles. *Q. J. R. Meteorol. Soc.*, **141**, 3018–3029, doi:10.1002/qj.2586. http://doi.wiley.com/10.1002/qj.2586.

Willoughby, H. E., 1979: Forced secondary circulations in hurricanes. *J. Geophys. Res.*, **84**, 3173, doi:10.1029/JC084iC06p03173. https://agupubs.onlinelibrary.wiley.com/doi/pdf/10.1029/JC084iC06p03173 (Accessed October 1, 2018).

Wu, L., H. Su, R. G. Fovell, T. J. Dunkerton, Z. Wang, and B. H. Kahn, 2015: Impact of environmental moisture on tropical cyclone intensification. *Atmos. Chem. Phys.*, **15**, 14041–14053, doi:10.5194/acp-15-14041-2015.

Zhang, D., and R. A. Anthes, 1982: A High-Resolution Model of the Planetary Boundary Layer—Sensitivity Tests and Comparisons with SESAME-79 Data. *J. Appl. Meteorol.*, **21**, 1594–1609, doi:10.1175/1520-0450(1982)021<1594:AHRMOT>2.0.CO;2. http://journals.ametsoc.org/doi/abs/10.1175/1520-0450(1982)021%3C1594:AHRMOT%3E2.0.CO;2.