An Ensemble Approach to Investigate Tropical Cyclone Intensification in Sheared Environments. Part I: Katia (2011)

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ABSTRACT

The mechanisms responsible for tropical cyclone (TC) intensification in the presence of moderate vertical shear magnitudes are not well understood. To investigate how TCs intensify in spite of moderate shear, this study employed a 96-member ensemble generated with the Advanced Hurricane Weather Research and Forecasting (AHW) Model. In this first part, AHW ensemble forecasts for TC Katia (2011) were evaluated when Katia was a weak tropical storm in an environment of 12 m s\(^{-1}\) easterly shear. The 5-day AHW forecasts for Katia were characterized by large variability in the intensity, presenting an opportunity to compare the underlying mechanisms between two subsets of members that predicted different intensity scenarios: intensification and weakening. The key difference between these two subsets was found in the lower-tropospheric moisture north of Katia (i.e., right-of-shear quadrant). With more water vapor in the lower troposphere, buoyant updrafts helped to moisten the midtroposphere and enhanced the likelihood of deep and organized convection in the subset that predicted intensification. This finding was validated with a vorticity budget, which showed that deep cyclonic vortex stretching and tilting contributed to spinning up the circulation after the midtroposphere had moistened. Sensitivity experiments, in which the initial conditions were perturbed, also demonstrated the importance of lower-tropospheric moisture, which suggests that moisture observations may help reduce uncertainty in forecasts of weak, sheared tropical storms.

1. Introduction

Even though vertical wind shear is typically detrimental for tropical cyclone (TC) development, TCs can form and intensify within moderate ambient shear values (≥5 m s\(^{-1}\)). In a climatological study of TC genesis and deep-layer (200–850 hPa) shear, Nolan and McGauley (2012) found that 5.0–8.75 m s\(^{-1}\) shear was more favorable for TC genesis than nearly zero shear. With a different approach, Hendricks et al. (2010) compared composites of intensifying and weakening TCs and found that even the intensifying TCs were surrounded by moderate shear of 6–10 m s\(^{-1}\). Several observational studies have also documented individual TCs that intensified in spite of moderate shear (e.g., Molinari et al. 2004, 2006; Shelton and Molinari 2009; Montgomery et al. 2010; Foerster et al. 2014). Despite these observations, a formal understanding of how TCs intensify amid moderate shear is limited and demands further investigation.

Moderate vertical wind shear affects the convective and kinematic structure of TCs (Corbosiero and Molinari 2002; Chen et al. 2006; Reasor et al. 2013; DeHart et al. 2014), which in turn can affect TC intensity. Previous studies that analyzed either cloud-to-ground lightning strikes (Corbosiero and Molinari 2002), satellite-derived rainfall (Chen et al. 2006; Hence and Houze 2011), or airborne Doppler radar observations (Reasor et al. 2013; DeHart et al. 2014) consistently found that convective activity preferentially occurs in the downshear half of sheared TCs. Along with the asymmetrically distributed moisture observed ...
convection, deep low-level inflow and midlevel ascent occurs downshear right, upper-level outflow primarily appears downshear left, and deep descent dominates the upshear quadrant (Reasor et al. 2013; DeHart et al. 2014). Because of this asymmetric nature of TCs in moderate shear, it is difficult to explain TC intensity changes on the basis of existing axisymmetric theory (Charney and Eliassen 1964; Emanuel 1986; Van Sang et al. 2008).

Previous studies have presented several pathways by which vertical wind shear can halt TC intensification (e.g., DeMaria 1996; Riemer et al. 2010; Tang and Emanuel 2010, 2012; Ge et al. 2013). Using a two-layer model to simulate the tilt of a TC vortex in moderate shear, DeMaria (1996) found a negative temperature anomaly above the low-level TC center. This temperature anomaly reduces the static stability above the low-level center, thus limiting convection within the TC. Using more sophisticated numerical simulations, Riemer et al. (2010) found that an effect of vertical wind shear is the reduction of boundary layer entropy by downdrafts. The reduced-entropy air, located left of the shear vector, suppresses convection within the eyewall and limits TC intensification. A similar finding was also documented by Tang and Emanuel (2012), who demonstrated that asymmetric fluxes of low-entropy air in the lower or midtroposphere reduce the conversion of available potential energy to kinetic energy in a TC. Additionally, Ge et al. (2013) found that the presence of dry air to the right of the shear vector can also be detrimental to TC intensity, given that cyclonic advection of dry air limits convection within the downshear quadrant.

The complexity of TCs and ambient shear suggests that the interaction between small-scale convective processes and the large-scale environment could be relevant to TC intensification. For example, previous studies have found that TC intensification in sheared environments can be triggered by the formation of a new TC vortex within the downshear convection (Molinari et al. 2004, 2006; Shelton and Molinari 2009). Molinari et al. (2004) suggested a two-stage intensification: first, the new vortex forms within the downshear convection, and, second, the TC intensifies via strong surface heat fluxes, as proposed by Emanuel (1986). However, not all TCs that undergo a “downshear reformation” continue to intensify after the new vortex forms (Molinari et al. 2006). It is possible that different environments or different TC structures result in different outcomes for the intensity evolution of a TC interacting with moderate shear.

While extensive literature exists on TCs and shear interactions, the intensification of weak, disorganized TCs (i.e., tropical depressions or storms) in moderate shear has not received much attention. To this end, the

![Fig. 1. Composite hodograph from the 96-member AHW ensemble forecasts initialized at 0000 UTC 30 Aug 2011. Dots represent area-averaged winds within a 500-km radius of Katia on isobaric surfaces extending from 850 to 200 hPa (every 50 hPa).](image)
presented in sections 2 and 3, respectively. The approach to diagnose the forecasts is discussed in section 4. Results are presented in sections 5–8, followed by the conclusions of this first part in section 9.

2. Experimental setup

High-resolution, full physics, ensemble forecasts were produced with the Advanced Hurricane Weather Research and Forecasting (WRF) Model (AHW) (Davis et al. 2008, 2010). The AHW Model was run twice a day in real time from 2009 to 2013 associated with the Hurricane Forecast Improvement Project (HFIP; Gall et al. 2013), and has been previously employed to investigate different aspects of TCs (Davis et al. 2010; Torn and Davis 2012; Galarneau and Davis 2013; Torn and Cook 2013; Brown and Hakim 2015). Experimental setup and data assimilation procedures are similar to those documented in previous studies (e.g., Torn 2010; Cavallo et al. 2013); therefore, only a brief summary is presented here.

The AHW ensemble prediction system was initialized from a cycling 96-member ensemble Kalman filter (EnKF) data assimilation system. At every analysis time, the only differences between ensemble members were small perturbations, consistent with analysis errors (Cavallo et al. 2013), to initial and boundary conditions. Each member had a fixed 36-km grid spacing domain, with an additional 12-km grid spacing domain centered on any disturbance or TC tracked by the National Hurricane Center. Observations were assimilated into both domains of all members in 6-hourly intervals from 0000 UTC 1 August 2011 to 1800 UTC 31 October 2011. Assimilated observations included ASOS, ACARS, ships, buoys, rawinsondes, atmospheric motion vectors (Velden et al. 2005), TC position and minimum sea level pressure (MSLP), and dropsonde data outside a radius of 200 km from the TCs (Cavallo et al. 2013). Following the assimilation, each member was advanced forward 6 h with the Advanced Research WRF (ARW) Model, version 3.1 (Skamarock et al. 2008), using the physics parameterizations listed in Table 1. Ensemble analyses from the cycling data assimilation were used to initialize 5-day forecasts with an additional 4-km grid spacing domain that followed the TC and did not employ a cumulus parameterization scheme.

3. Synopsis of Katia (2011) and the AHW forecasts

Avila and Stewart (2013) summarized the evolution of Katia as follows. The genesis of Katia was associated with the interaction between an African easterly wave and accompanying low pressure system emerging off Africa on 27 August 2011. Convection near the system gained organization, leading to the formation of a tropical depression at 0600 UTC 29 August. Despite being influenced by moderate-to-strong deep-layer shear of 12.5–15 m s$^{-1}$, the tropical depression continued intensifying until becoming a tropical storm at 0000 UTC 30 August and a hurricane at 0000 UTC 31 August. As Katia continued moving westward over the eastern Atlantic, intensification ceased, and Katia remained at $\sim$33 m s$^{-1}$ intensity for 3 days. Following this period, Katia rapidly intensified and reached a peak intensity of 62 m s$^{-1}$ at 0000 UTC 6 September. After encountering cold SSTs, Katia weakened and transitioned into a powerful extratropical system on 10 September.

*AHW ensemble forecasts*

This study focuses on 5-day forecasts initialized at 0000 UTC 30 August 2011, when Katia was affected by 200–850-hPa easterly shear of approximately 12 m s$^{-1}$, as well as by 6 m s$^{-1}$ shear between 600 and 850 hPa (Fig. 1). As shown in Fig. 2a, the track forecasts captured the observed motion of Katia; the observed track of Katia fell within the AHW ensemble spread during most of the forecast period, except for the last 24 h. Despite the good agreement in track forecasts, the ensemble was characterized by large variability in MSLP forecasts (Fig. 2b). The largest separation between ensemble members happened after 48 h, when the ensemble standard deviation went from 3 hPa at 48 h to 16 hPa at 126 h. Ensemble forecasts initialized within 24 h before and after 0000 UTC 30 August 2011 were also characterized by remarkably large spread (not shown); thus, the forecasts studied here are not unique. This large spread in the ensemble forecasts provided motivation to investigate the underlying physical processes that led to intensification in some members but weakening or steady state in other members. It should be noted that all ensemble members underestimated the MSLP between 0 and 48 h, but several members captured the intensification after 48 h. Nonetheless, the ensemble forecasts provide valuable

<table>
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<th>Table 1. List of physics parameterizations used with the AHW Model.</th>
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<td>Parameterization</td>
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<tr>
<td>Kain–Fritsch cumulus convection</td>
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<tr>
<td>Rapid Radiative Transfer longwave radiation</td>
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<tr>
<td>Dudhia shortwave scheme</td>
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<td>WRF single-moment 6-class (WSM6) microphysics</td>
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<td>Yonsei University (YSU) planetary boundary layer</td>
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<td>Similarity theory land surface model$^a$</td>
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$^a$ Includes the updated enthalpy and momentum drag formulations described in Davis et al. (2010).
information about the physical processes behind different intensity evolutions of sheared TCs.

4. AHW analysis methodology
   a. Forecasts diagnoses approach

   To investigate the processes leading to different intensity evolutions in the AHW ensemble, two subsets of 12 members each were identified: one subset with the members that predicted the lowest mean 0–126-h MSLP (herein called strong members) and another subset with the members that predicted the highest mean 0–126-h MSLP (herein called weak members). The rest of this manuscript will explore the differences between these two subsets. Rather than considering individual members, composites of these two subsets were evaluated at each forecast time. For reference, the composite track and intensity forecasts are shown in Fig. 2. This composite approach has been successfully implemented in previous observational (e.g., Rogers et al. 2013) and modeling (e.g., Munsell et al. 2013) studies to compare intensifying and weakening (or steady state) TCs. Using the composite method in this study allows us to compare similarities and differences between ensemble members, while also permitting the results to be tested for statistical significance.

   Composite differences between strong and weak members were tested for statistical significance at the 95% confidence interval using a bootstrap resampling approach as follows. First, two subsets of 12 members each were randomly drawn from the 96-member ensemble. Second, the difference between the composites of randomly selected subsets was computed. These two steps were repeated 10000 times to create a distribution, which was then used to test the null hypothesis that the difference between strong and weak members was identical to the difference between two subsets of randomly selected members. This approach allowed for a statistical significance assessment without assuming that the ensemble matched a specific probability distribution.

Fig. 2. The AHW ensemble (a) track and (b) intensity forecasts initialized at 0000 UTC 30 Aug 2011. The gray, red, and blue lines depict the individual ensemble members, the composite of strong members, and the composite of weak members, respectively. National Hurricane Center best tracks are shown with the black lines. The squares in (a) show 6-hourly storm positions.
b. Water vapor budget

Given the importance of moist convective processes on TC intensification, a water vapor budget was considered to compare strong and weak members. The budget was computed using the Boussinesq water vapor equation:

\[
\frac{\partial (\rho q)}{\partial t} = -\mathbf{V}_h \cdot (\rho q \mathbf{V}_h) - \frac{\partial (\rho q w)}{\partial z} - C + E + B,
\]

where \(q\) is the water vapor mixing ratio, \(\rho\) is the dry air density, \(\mathbf{V}_h\) is the horizontal wind, \(w\) is the vertical velocity, \(C\) is condensation, \(E\) is evaporation, and \(B\) is the contribution from the planetary boundary layer parameterization scheme. The first and second terms on the right-hand side of Eq. (1) represent the horizontal and vertical water vapor flux convergence, respectively. The third and fourth terms combined together yield the water vapor tendency due to the net condensation within the microphysics parameterization scheme. Integrating over \(A\), the area of a circular region centered on the TC, and over a time period between \(t_0\) and \(t_1\), Eq. (1) becomes:

\[
\rho \tilde{q}_v(t_1) - \rho \tilde{q}_v(t_0) = -\frac{1}{A} \int_{t_0}^{t_1} \int_A \mathbf{V}_h \cdot (\rho q \mathbf{V}_h) dA dt \\
- \frac{1}{A} \int_{t_0}^{t_1} \int_A \frac{\partial (\rho q w)}{\partial z} dA dt \\
- \int_{t_0}^{t_1} (\tilde{C} - \tilde{E}) dt + \int_{t_0}^{t_1} \tilde{B} dt,
\]

where the tilde is used to denote area-averaged quantities. Equation (2) helps determining the relative contributions from different processes to the water vapor change between forecast times \(t_0\) and \(t_1\).

c. Vorticity budget

A vorticity budget was also employed to evaluate the dynamical processes behind the different intensity changes within the AHW ensemble. In this study, the vorticity budget is the same as in Davis and Galarneau (2009), where an expression for the temporal change of the circulation \(C\) was presented as follows:

\[
\frac{\partial C}{\partial t} = -\nabla \tilde{\delta} A - \int_A \tilde{\eta} \mathbf{V}_h \cdot \hat{n} dl + \oint_A \left( \mathbf{\tilde{\omega}} \times \frac{\partial \mathbf{V}_h}{\partial p} \right) \cdot \hat{n} dl \\
+ \oint_A (\mathbf{\tilde{k}} \times \mathbf{F}) \cdot \hat{n} dl,
\]

where \(\eta\) is the absolute vorticity, \(\tilde{\delta}\) is the mean divergence over an area \(A\), \(\mathbf{V}_h\) is the storm-relative horizontal wind vector, \(\omega\) is the vertical velocity on isobaric coordinates, and \(\mathbf{F}\) is the friction vector. Equation (3) contains closed integrals along a box centered on the TC; thus, \(\hat{n}\) is a unit vector perpendicular to the edges of the box, \(dl\) is the differential along the perimeter of the box, the overbar symbol represents the mean along the edge of the box, and the prime symbol represents the perturbation from that mean. Partitioning the horizontal wind components into means and perturbations from the means conveniently separates the contributions to the local change in circulation into the following terms (from left to right): stretching, eddy vorticity flux, tilting, and friction (Davis and Galarneau 2009). Similar to the water vapor budget, Eq. (3) is integrated from \(t_0\) to \(t_1\) and normalized by \(A\):

\[
\tilde{\zeta}(t_1) - \tilde{\zeta}(t_0) = -\int_{t_0}^{t_1} \eta \tilde{\delta} dt - \frac{1}{A} \int_{t_0}^{t_1} \oint_A \tilde{\eta} \mathbf{V}_h \cdot \hat{n} dl dt \\
+ \frac{1}{A} \int_{t_0}^{t_1} \oint_A \left( \mathbf{\tilde{k}} \times \frac{\partial \mathbf{V}_h}{\partial p} \right) \cdot \hat{n} dl dt \\
+ \frac{1}{A} \int_{t_0}^{t_1} \oint_A (\mathbf{\tilde{k}} \times \mathbf{F}) \cdot \hat{n} dl dt,
\]

where circulation has been replaced by the area-averaged vorticity. Equation (4) provides a convenient tool to quantify the different contributions to intensity changes as measured by the area-averaged vorticity.

5. Diagnoses of TC Katia forecasts

a. Comparison of environments

Since previous studies have suggested that TC intensity is highly influenced by the environment (e.g., Gray 1968; Merrill 1988; Emanuel et al. 2004; Hendricks et al. 2010; Torn and Cook 2013), the role of environmental differences on subsequent intensity is explored first. Area-averaged quantities were computed with the aim of obtaining a representation of the environment while removing most of the contributions from the TC itself. Figure 3 shows time series of the 200–850-hPa vertical wind shear\(^1\) and precipitable water (PW) averaged within 500 km of Katia. Other environmental quantities were considered (e.g., SST, ocean heat content, and upper-level divergence), but no statistical significance was found.

As seen in Fig. 3, the environment of strong members is slightly more favorable for intensification up to 72 h. Even though all composites are characterized by 12 m s\(^{-1}\) shear during the first 12 h, strong members have statistically significantly smaller shear magnitude than weak members.

\(^1\) Here, shear was calculated by first removing the rotational and divergent components of the TC vortex using the method of Galarneau and Davis (2013), then averaging the remaining winds within a 500-km radius.
members during 12–36 h (Fig. 3a). During this period, shear values decrease as Katia moves away from the tropical easterly jet. Shear values increase again after 48 h, when Katia approaches an upper-tropospheric trough. Strong members have significantly greater shear after 76 h, albeit at a time when strong members have deepened 35 hPa more than weak members (Fig. 2b). On the other hand, the PW shows one of the clearest and most consistent distinctions between the subsets (Fig. 3b). Strong members have PW exceeding 56 mm, while weak members have less than 56 mm between 0 and 84 h. All members see a 2–5-mm reduction in PW as Katia moves away from the ITCZ, but strong members move farther north than weak members (Fig. 2a) and have less PW after 96 h.

These findings are consistent with previous studies that compared intensifying and weakening (or steady state) TCs (e.g., Hendricks et al. 2010; Wu et al. 2012): weaker shear and more near-storm moisture aids intensification. Nonetheless, it is intriguing that strong members are able to intensify amid initially moderate shear. The greater environmental moisture of strong members could mitigate entrainment and also favor more convective activity. Given the clear distinction in the PW of strong and weak members, it is worth further investigating the location and timing of moisture differences.

b. Storm-relative moisture differences

Instead of comparing composites of the convective evolution of both subsets, one member from each subset is examined in order to retain the details of the TC structure (Fig. 4). The members shown in Fig. 4 have the closest MSLP forecast to the composite of their respective subsets, based on a root-mean-square error test. It should be noted that the evolutions of these two members are similar to the rest of the members of each subset; thus, the results are representative of each subset.

As shown in Fig. 4, convective activity related to Katia evolves differently in the strong and weak members. The 6-h simulated composite reflectivity of both members shows regions of active convection exceeding 50 dBZ to the southwest of Katia, albeit more abundant in the strong member (Figs. 4a,b). Given the easterly shear, the highest reflectivity is located within the downshear quadrant of Katia. As the forecast evolves, the most active convection rotates around Katia, focusing to the north and west (i.e., downshear and right of shear) of Katia by 24 h (Figs. 4c,d). A banding feature is evident north of Katia in the strong member and persists throughout 48 h, when the strong member has noticeable eyewall and rainbands (Figs. 4e,f). In contrast, the weak member is characterized by disorganized convection confined to the downshear quadrant. By this time, the shear of the strong member has changed from easterly to westerly shear. The weak member also encounters westerly shear after 48 h, but this member remains weak and disorganized throughout the forecast (not shown). The strong member, however, becomes more symmetric and continues intensifying beyond 48 h.

Based on these different evolutions during the first 48 h, a plausible hypothesis is that the differences in the convective evolution of strong and weak members could be related to moisture. Storm-centered composite differences on a latitude–longitude grid are considered to test this hypothesis (Fig. 5). Composite differences are normalized by the standard deviation of the full ensemble to facilitate the comparison between different variables at different vertical levels.

From the onset, strong members have 0.8–2.0 standard deviations (σ) more PW than weak members (Fig. 5a). The largest storm-centered differences are not located at the center of Katia; instead, these differences are 2°–4° (approximately 200–400 km) away from...
Fig. 4. Composite reflectivity over the lowest 3-km layer (shading; every 5 dBZ) of (left) a strong member and (right) a weak member of the ensemble forecasts of Katia. The inset in the lower-right corner of each panel shows the 200–850-hPa shear vector direction and magnitude. Three forecast times are shown: (a),(b) 6; (c),(d) 24; and (e),(f) 48 h.
from the center. Since PW is a vertically integrated quantity, moisture differences are further investigated at individual vertical levels. The largest moisture differences located 3°–4° north of Katia at 0 h are most prominent at 950 hPa (Fig. 5b), whereas the largest differences located 1°–2° west of Katia are most prominent at 500 hPa (Fig. 5c). These differences suggest there could be two different origins of more

FIG. 5. Latitude–longitude analysis of storm-centered composites relative to the 850-hPa vortex center of the ensemble forecasts of Katia. Shown are the standardized composite differences between the mean of strong members and the mean of weak members (shading, every 0.4σ) of (a),(d),(g) precipitable water vapor; (b),(e),(h) 950-hPa water vapor mixing ratio; and (c),(f),(i) 500-hPa water vapor mixing ratio. The ensemble mean of each quantity is shown by the black contour lines every 2 mm in (a),(d), and (g); every 1 g kg⁻¹ in (b),(e), and (h); and every 0.5 g kg⁻¹ in (c),(f), and (i). Statistically significant differences are denoted by the stippling pattern. Three forecast times are shown: (top) 0, (middle) 24, and (bottom) 48 h.
moisture in strong members: north of Katia at low levels and southwest of Katia at midlevels.

As the forecast progresses, moisture differences amplify in magnitude and expand in spatial coverage (Figs. 5d–f). Storm-relative composite differences at 24 h indicate more moisture in strong members north, west, and south of Katia. There is a region of higher PW that wraps cyclonically around Katia, originating 5° east and finishing within the center (Fig. 5d). This feature is most evident at low levels, indicated by a greater water vapor mixing ratio at 950 hPa (Fig. 5e). By contrast, the largest differences at 500 hPa are located south of Katia (Fig. 5f). Later on, the 48-h forecast shows strong correspondence between PW differences and 500-hPa mixing ratio differences, where there is 1.2–2.0σ more water vapor in strong members (Figs. 5g–i). With more midlevel moisture, strong members are more likely to have sustained, deep moist convection, as noted in the composite reflectivity at 48 h (Figs. 4e,f). Consequently, near-storm moisture differences continue to amplify beyond 48 h, especially at 500 hPa and above (not shown).

c. Shear-relative differences

This subsection will further examine the vertical location of moisture differences with respect to the shear vector because of the documented effect of shear on the distribution of convection (e.g., Corbosiero and Molinari 2002; Chen et al. 2006; DeHart et al. 2014). To this end, the storm is decomposed into the following shear-relative quadrants: downshear and upshear quadrants, which are defined ±45° of the shear vector, and right-of-shear and left-of-shear quadrants, which are defined ±45° perpendicular to the shear vector. Moist static energy (MSE), which is akin to equivalent potential temperature, is used in the shear-relative composites, as Raymond and Sessions (2007) demonstrated that variations in moisture can change the MSE and vertical mass flux profile. It should be noted that, even though MSE also depends on temperature, the composite differences to be discussed below are mainly attributed to the water vapor component of MSE.

Strong members have statistically significant higher MSE in all quadrants beginning at 0 h (Figs. 6a,d). The largest differences (2.0σ) in the lower troposphere appear in the right-of-shear quadrant, which corresponds to the differences north of Katia in Fig. 5b. A broad region of 2.0σ MSE differences appears in the downshear quadrant, where strong members have greater MSE from the surface up to 7-km height. The ensemble-mean structure suggests that the downshear differences are located within or close to a midlevel MSE minimum, which hints at less potential for entrainment in strong members. The upshear and left-of-shear quadrants are also characterized by greater MSE, particularly above 5 km. These results suggest that strong members have greater potential to initiate moist convection in the lower troposphere of the right-of-shear quadrant and enhance the convection as the parcels travel cyclonically into the downshear and left-of-shear quadrants.

As the forecast advances, the largest MSE differences between strong and weak members shift from the lower troposphere to the mid- and upper troposphere. The 24-h forecast indicates that strong members have more than 2.0σ greater MSE between 8- and 12-km height of the downshear and left-of-shear quadrant (Figs. 6b,e). Additionally, all quadrants show up to 1.5–2.0σ greater MSE in the midtroposphere of strong members. These results suggest strong members could have greater and deeper moist convection, particularly between the downshear and left-of-shear quadrants. By 48 h, this pattern appears more uniform as strong members have significantly greater MSE above 4 km of all quadrants (Figs. 6c,f). The largest differences (2.0–2.5σ) at this lead time are found within 100 km of the center, implying less vertical gradient of MSE and more potential for deep moist convection within the inner core of Katia. After 48 h, strong members continue intensifying with greater moisture and also develop a warm core, resulting in greater MSE of all quadrants (not shown).

6. Role of tropospheric moisture

The storm-relative composites suggest that the location of enhanced moisture is important for the intensification of strong members. Some of the largest differences initially appear 200–400 km away from the TC center; thus, it is important to understand how and why the enhanced moisture at that location affects the inner vortex of Katia. To investigate the relevance of the location of moisture differences, parcel trajectories are analyzed for a representative member of the strong subset. Approximately 170 parcels were released forward within the region of 950-hPa water vapor differences north of Katia (i.e., right of shear). Parcel trajectories were integrated forward using the Read/Interpolate/Plot (RIP) program with 6-min output from 0 to 18 h.

Parcels released within the 950-hPa water vapor differences follow a wide variety of trajectories (Fig. 7a). Most parcels that originate within 2° (≈200 km) north circulate around Katia several times, whereas most parcels that originate between 2° and 4° (≈200–400 km) only circulate around the western half and into the inner 1° (≈100 km) of Katia. These different trajectories also

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4 RIP is available online at [http://www2.mmm.ucar.edu/wrf/users/docs/ripug.htm](http://www2.mmm.ucar.edu/wrf/users/docs/ripug.htm).
result in different distributions of final pressure and final relative humidity (RH). While the vast majority of parcels remain between 800 and 1000 hPa, more than 10% of the parcels that originate within the innermost 2° ascend into the upper troposphere (Fig. 7b). More than 50% of the parcels become near-saturated regardless of their initial location with respect to Katia, but a significant number of parcels have RH below 50% at 18 h (Fig. 7c). These low RH values are associated with parcels that ascend into the upper troposphere, where phase changes reduce the RH with respect to liquid water.

A comparison of the final (Fig. 7c) and initial (Fig. 7d) RH distributions shows that most parcels go from 70%–80% RH at 0 h to 80%–100% RH at 18 h, especially those parcels that originate more than 2° away from Katia. Enhanced moisture north of Katia, or right of shear, is thus important, because those parcels become near saturated as they circulate around and into Katia, passing through the left-of-shear quadrant, where dry air is fluxed downward into the boundary layer of sheared TCs (Riemer et al. 2010). Moistening along the parcel trajectories is thus hypothesized to suppress the drying effects of convective downdrafts.

To further assess the relevance of water vapor differences, a quantitative approach is taken with the water vapor budget from Eq. (2). This budget was calculated using 3-min output from the AHW forecasts of strong and weak members. All terms were computed within a 200-km radius because the moisture differences were most prominent at or beyond that distance from Katia. Two periods were selected for the calculation: 1) from 12 to 24 h, when strong members have more water vapor in the lower troposphere and 2) from 36 to 48 h, when strong members have more water vapor in the midtroposphere. Results from the budget calculations, averaged for all members in each subset, are shown in Fig. 8. It should be noted that even

Fig. 6. Shear-relative composites of MSE normalized by $c_p$ as a function of height and radius at (a),(d) 0; (b),(e) 24; and (c),(f) 48 h. Shown are the standardized composite differences between the mean of strong members and the mean of weak members (shading, every 0.5σ). The ensemble mean (black contours; every 1 K) and statistical significance (stippling) are shown as in Fig. 5.
with 3-min output, the summation of all terms on the right-hand side of Eq. (2) does not match exactly the water vapor change given by the left-hand side of Eq. (2). These discrepancies are most certainly related to the temporal and spatial resolution near small-scale processes in the boundary layer or within convection.

During the 12–24-h period, both subsets are characterized by moistening below 3.5 km and drying above 3.5 km (Fig. 8a). Strong members have smaller water vapor changes than weak members, which results from large cancellations between individual tendencies (Fig. 8b). An evaluation of the individual tendencies shows that strong members have stronger horizontal water vapor flux convergence below 3 km, weaker horizontal water vapor flux divergence above 3 km, stronger vertical water vapor flux divergence below 2 km, stronger vertical water vapor flux convergence above 3.5 km, and greater condensation above 1 km. These results imply that the horizontal winds add more water vapor to the lower troposphere of strong members, where the vertical wind moves the water vapor up to the midtroposphere. Even though the cancellation between terms results in smaller water vapor changes in strong members, this subset has more vigorous and deeper convection.

During the 36–48-h period, the summation of the tendencies shows moistening below 1 km and drying above 1 km (Fig. 8c). However, the actual water vapor change (solid lines in Fig. 8c) shows mostly moistening in strong members and drying in weak members. This discrepancy between the summation of terms and the actual water vapor change could be related to an over-estimation of the net condensation term due to the presence of vigorous convection. Evaluating the individual tendencies shows that both subsets have less horizontal flux divergence than during the 12–24-h period, but strong members have greater horizontal water vapor...
flux convergence and vertical water vapor flux divergence in the lowest 1.5 km (Fig. 8d). Above 1.5 km, strong members have larger vertical water vapor flux convergence and more condensation. These results indicate that strong members are importing more water vapor in the lower troposphere, which supports more condensation and deeper convection near the center of Katia.

During both time periods considered, strong members are characterized by greater horizontal and vertical water vapor flux convergence within and above the boundary layer, respectively. This pattern is indicative of stronger or more abundant updrafts in strong members, which is also consistent with the low pressure and RH noted in the trajectory analysis. Since the storm-centered composite differences indicated that strong members have more low-level moisture than weak members, it is possible that the deep updrafts result from enhanced buoyancy. This mechanism would be consistent with the hypothesis posed by Dolling and Barnes (2012) that increased buoyancy, instability, and vortex stretching aid the intensification of tropical storms. This possibility is explored by evaluating the distribution of surface-based CAPE\(^5\) and convective\(^6\) vertical velocities between 500 and 850 hPa as proxies for buoyant convection. Only a strong and a weak member are considered because composites would remove the details.

\(^5\) CAPE values presented here are likely overestimated because water loading and entrainment are not considered in the calculation.

\(^6\) Convective vertical motion implies that we only considered grid points with cloud mixing ratio greater than \(1 \times 10^{-5}\) kg kg\(^{-1}\).

FIG. 8. Vertical profiles of water vapor budgets during (a),(b) 12–24 and (c),(d) 36–48 h. (left) The change in water vapor (solid lines) and the sum of all terms of the water vapor equation (dashed lines). (right) The contributions to the water vapor budget: horizontal water vapor flux convergence (solid lines), vertical water vapor flux convergence (short-dashed lines), and the combination of evaporation minus condensation with the boundary layer term (long-dashed lines). In all panels, the red (blue) lines represent the mean of strong (weak) members. Notice the irregularly spaced abscissa.
As illustrated in Fig. 9, the strong member has larger CAPE only during the period of largest low-level water vapor differences between strong and weak members. The 24-h forecast shows regions where CAPE exceeds 2000 J kg$^{-1}$ north and west of Katia in the strong member (Fig. 9a), but CAPE values of the weak member are less than 1500 J kg$^{-1}$ (Fig. 9b). Along with higher CAPE, the strong member also has more convective updrafts in a banding feature north, west, and south of Katia. This banding feature was noted before as a region of more 950-hPa water vapor in strong members. To prove that the strong member has larger CAPE and more updrafts due to greater moisture in the lower troposphere, Fig. 9c shows the CAPE obtained using the temperature from the weak member and the water vapor mixing ratio from the strong member. Figures 9a and 9c are practically similar everywhere, which demonstrates that buoyancy differences between strong and weak members arise from the near-surface moisture rather than the temperature profile. CAPE values of both members increase beyond 2500 J kg$^{-1}$ by 48 h (Figs. 9d,e), when the largest differences between strong and weak members are no longer in the lower troposphere (cf. Figs. 5g–i and 6c,f). Even with comparable near-surface moisture and buoyancy, the strong member continues to have more updrafts, while the weak member has more downdrafts near the center of Katia.

While Fig. 9 suggests a higher occurrence of buoyancy-driven convective updrafts (downdrafts) near Katia in strong (weak) members, it is difficult to assess the actual magnitude and number of updrafts and downdrafts. To assess if and when strong members have more convective updrafts, probability distributions of 500–850-hPa layer-averaged vertical velocity within a 200-km radius are evaluated using hourly output from all strong and weak members. The distributions, shown in Fig. 10, were constructed by counting the number of convective vertical velocities into geometrically spaced bins and then normalizing with respect to the total number of data points used at each hour. Using this method demonstrates that...
strong members have stronger updrafts than weak members; while the peak (14%–16% chances) of strong members is between 8 and 16 cm s\(^{-1}\) (Fig. 10a) throughout the forecast, the peak (18%–20% chances) of weak members is centered on 0 cm s\(^{-1}\) (Fig. 10b). Taking the difference between the two distributions illustrates that strong members are more likely to have updrafts with magnitudes greater than 4 cm s\(^{-1}\), whereas weak members mostly have weak updrafts and downdrafts between -4 and 4 cm s\(^{-1}\) (Fig. 10c). These differences amplify over time, highlighting the relevance of the initial lower-tropospheric moisture differences in initiating more buoyant convection in strong members.

As strong members are able to sustain stronger updrafts, the latent heat released by convection is expected to be greater and aid more deepening in this subset. However, the latent heat is nonuniformly distributed because of the asymmetric nature of Katia. To illustrate this, Fig. 11 shows the condensational heating partitioned into the shear-relative quadrants and volume integrated from the lowest to the highest model level within a 200-km radius of Katia. The largest values of the integrated heating, as well as the largest differences between strong and weak members, appear first in the downshear quadrant (Fig. 11a). Although there is a reduction in the downshear-integrated heating between 30 and 36 h, strong members continue to have more latent heat release, albeit in the left-of-shear quadrant (Fig. 11b). After 36 h, the integrated heating of strong members also increases in the upshear quadrant (Fig. 11c), followed by an increase in right-of-shear heating around 40 h (Fig. 11d). These results project onto Fig. 10 as a continuous peak in convective updrafts activity, because strong members have more convection in one or more quadrants at all times. The ability of strong members to sustain deep convection and latent heat release could be accompanied by vortex stretching, which will be investigated in the next section.

7. Quantification of the intensity changes

In this section, the vorticity budget is evaluated to discern the different mechanisms aiding or inhibiting the spinup of Katia, and to compare those mechanisms with the results from the previous sections. Equation (4) was evaluated for the same two periods considered for the water vapor budget, with a 300 km \times 300 km box during 12–24 h and a 200 km \times 200 km box during 36–48 h. These boxes enclosed the inner-core circulation of Katia as measured by the radius of maximum wind, which went from 150 km during 12–24 h to 100 km during 36–48 h. Similar to Davis and Galarneau (2009) and Galarneau et al. (2013), the center point of the box was moved ±20 km from the TC center in 4 km increments to diminish the biasing effects of deep convective activity at

Fig. 10. Time-bins analysis of the distribution of layer-averaged vertical velocity between 500 and 850 hPa and within a 200-km radius of Katia. (a), (b) The normalized likelihood of vertical velocities in each bin (shading; every 2%) corresponding to strong and weak members, respectively. (c) The difference between the distributions of strong and weak members (shading; every 1%).
the edge of the box. This method was repeated for strong and weak members, and the results will be presented as area-averaged means of each subset. Hourly output was sufficient to obtain a budget that is similar in shape and close in magnitude to the actual area-averaged vorticity change during the two periods considered (Fig. 12).

During the 12–24 h period, the area-averaged vorticity of strong members increases at all levels, whereas the area-averaged vorticity of weak members only increases between 900 and 825 hPa (Fig. 12a). An evaluation of the individual tendencies reveals that strong members have greater vortex stretching at all levels considered (Fig. 12b). The contribution of the vortex stretching in the budget of strong members is positive up to just above 725 hPa; thus, other mechanisms must be contributing to the increased area-averaged vorticity in the midtroposphere. The eddy vorticity flux of strong members is small and negative at all levels considered, but the tilting term of strong members is dominantly positive above 750 hPa. Weak members, however, have more negative stretching and less tilting than strong members, resulting in a smaller change of area-averaged vorticity. The key differences between the subsets are associated with the greater number of updrafts in strong members, which contributes to both low-level convergence of absolute vorticity and tilting of horizontal vorticity. The tilting term can be further interpreted as the correlation between vertical velocity and the component of horizontal vorticity normal to the boundary of the TC. This correlation is greatest in the downshear quadrant of strong members, where shear-forced updrafts generate cyclonic relative vorticity inside the box by tilting the outward-pointing horizontal vorticity that results from storm-scale tangential winds decreasing with height (not shown).

Substantial changes to the area-averaged vorticity profiles occur during the 36–48-h period: the vorticity of strong members strengthens, but the vorticity of weak members weakens at all levels considered (Fig. 12c). The summation of all tendencies is close to the actual change in the area-averaged vorticity of both members but results in a different sign in weak members. This discrepancy is likely attributed to diffusion, numerical errors, and the temporal resolution of the output. Nevertheless, the summation of accumulated tendencies is considerably smaller for weak members than for strong members, which gives confidence on the different contributions to the vorticity budget. Contrary to the 12–24-h period, strong members now have positive contributions from vortex stretching at all levels (Fig. 12d). The eddy vorticity term contributes a positive vorticity change between 900–700 hPa but a negative change above 700 hPa. However, tilting of horizontal vorticity dominates the positive vorticity change from 850 to 600 hPa and above 500 hPa as deep updrafts are collocated with outward-pointing horizontal vorticity. Weak members have a different scenario, with weaker tilting than strong members, as well as near-zero

![Fig. 11. Time series of the volume-integrated condensational heating decomposed by shear quadrants as follows: (a) upshear, (b) downshear, (c) left of shear, and (d) right of shear. Black, red, and blue lines depict the mean of the 96-member ensemble, the mean of strong members, and the mean of weak members, respectively. Dots represent statistically significant differences between strong and weak members at the 95% confidence interval.](image-url)
and even negative vortex stretching and eddy vorticity flux (Fig. 12d).

The vorticity budget presented here suggests a two-stage intensification process. During the first stage, convergence over a shallow layer contributes to cyclonic vortex stretching in the lower troposphere. Tilting of the horizontal vorticity contributes to cyclonic vorticity enhancement in the midtroposphere, but it is offset by negative vortex stretching. This process is consistent with the results of the previous section, in which strong members had more water vapor in the lower troposphere, which aided buoyant convection over a shallow layer. During the second stage, deeper and stronger updrafts contribute to cyclonic vortex stretching over a deeper layer. Greater convection downshear and left of shear results in positive tilting, thus increasing the area-averaged vorticity even more in the midtroposphere. This process coincides with the timing when strong members have more moisture at mid- and upper levels, which likely favored deeper and more vigorous convection. After this stage, a continuous feedback is established between the strong surface winds, surface fluxes, and convection such that strong members continue intensifying throughout the forecast.

8. Sensitivity experiments

The previously discussed results suggest that uncertainty in the analyzed lower-tropospheric moisture contributes to uncertainty in the intensity forecasts. A
series of sensitivity experiments was conducted to investigate the changes in the intensity forecasts corresponding to perturbed initial conditions (ICs) in response to changes in the lower-tropospheric moisture. Similar to the methods of Torn and Hakim (2009), Torn (2010), Torn and Cook (2013), and Brown and Hakim (2015), the ICs for the 96-member ensemble were perturbed via:

$$\mathbf{x}^p_i = \mathbf{x}^c_i + \frac{\partial \mathbf{x}^c_i}{\partial J_{IC}} \alpha,$$

where $\mathbf{x}^p_i$ and $\mathbf{x}^c_i$ are the ensemble of perturbed and control ICs, respectively, of the $i$th state variable. The second term on the right-hand side of Eq. (5) is the perturbation to the control ICs and is given by

$$\frac{\partial \mathbf{x}^c_i}{\partial J_{IC}} \alpha = \frac{\text{cov} (\mathbf{x}^c_i, J_{IC})}{\text{var} (J_{IC})} \alpha,$$

where $J_{IC}$ represents the ensemble estimate of the variable to be changed (herein referred to as the IC metric), and $\alpha$ represents the amplitude of the change. This method is akin to a data assimilation, except that the gain is given by $\partial \mathbf{x}^c_i / \partial J_{IC}$, and the increment is prescribed in $\alpha$. After perturbing the ICs, AHW forecasts were integrated forward up to 72 h.

The sensitivity experiments assumed that observations of the 950-hPa MSE\(^7\) north of Katia were available. This IC metric (950-hPa MSE) was averaged only within grid points where the 0-h water vapor difference between strong and weak members was statistically significant (cf. Fig. 5b). Three different experiments were performed with this IC metric. The first experiment (named ALL) consisted of IC perturbations to all state variables via Eq. (5), with different $\alpha$ values to represent different magnitudes of MSE at 950 hPa. The second experiment (named MOIST) consisted of IC perturbations only to water mixing ratios, while the third experiment (named DRY) contained IC perturbations to all variables, except for the water mixing ratios. The purpose of these additional experiments was to discern if the moisture perturbations were linked to the initial strength of the vortex. In all experiments, $\alpha$ values ranged between $2 \times 10^{-5}$ and $1 \times 10^{-5}$, where $\sigma$ is the ensemble standard deviation of the IC metric.

Perturbing the initial conditions of all members based on $1\sigma$ amplitude results in important changes to the ICs of the AHW ensembles. Figure 13 shows the difference between the ensemble mean of the ALL sensitivity experiment and the control simulation. (a) The 950-hPa water vapor mixing ratio at 0 h, and the 850-hPa area-averaged vorticity (scaled by $1 \times 10^{-5}$) at (b) 0 and (c) 72 h.

\(^7\)MSE is used here instead of water vapor mixing ratio because both variables yield very similar composite differences between strong and weak members.
ALL experiment than in the control simulation. The 72-h forecast of the sensitivity experiment portrays a stronger circulation between 1° south and 2° north of Katia, with the largest differences collocated within the vortex itself (Fig. 13c). This result hints that the intensification of Katia could be influenced by either the initial near-storm moisture or the strength of the lower-tropospheric circulation. Furthermore, the temporal evolution of the circulation within 100 km of Katia illustrates that the ALL experiment evolves from a stronger circulation at low levels to a stronger circulation throughout 10-km height after approximately 12 h (Fig. 14a). The circulation differences appear after the sensitivity experiments have more condensational heating (Fig. 15a), which suggests that increased low-level moisture enhances the buoyancy-driven updrafts that strengthen Katia via vortex stretching and tilting of horizontal vorticity.

Perturbing only the water mixing ratios because of an increase of 1σ in the 950-hPa MSE has a similar effect as perturbing all state variables. Figure 14b demonstrates that the ensemble mean of the MOIST experiment also evolves into a stronger TC by 72 h. There is also greater condensational heating in this experiment as compared to the control, which is consistent with the aforementioned results on the role of the lower-tropospheric moisture (Fig. 15b). However, if only the kinematic field is perturbed based on the same metric and amplitude, the ensemble-mean changes are smaller than when perturbing the water mixing ratios. As seen in Fig. 14c, the difference in circulation between the DRY experiment and the control is almost zero through 12 h, after which time small changes appear. The DRY experiment also has less condensational heating during certain periods of time (e.g., 0–12 h and 30–42 h) (Fig. 15c), indicating that there is less convection than in the control simulation even though the ICs are perturbed to yield an initially stronger TC.

The sensitivity experiments based on the initial 950-hPa MSE demonstrate that moisture differences between strong and weak members arise from the near-storm moisture and not from differences in the lower-tropospheric circulation. Only one set of experiments with a perturbation magnitude of 1σ has been presented thus far, but the same result holds true for different perturbation magnitudes. Figure 16 summarizes the results from all sets of experiments with different σ values by comparing the predicted and actual 72-h MSLP forecast changes. Here, “predicted” refers to the value given by Eq. (6) (except χ now represents the 72-h MSLP forecast of the control) and “actual” refers to the difference between the 72-h MSLP forecast from the experiments and the 72-h MSLP forecast from the control simulations.

Comparing the predicted and actual 72-h MSLP forecast changes reveals some important differences between the experiments (Fig. 16). Generally, there is good agreement between the actual and predicted MSLP changes in the ALL and MOIST experiments but not a good agreement in the DRY experiments. While the largest actual changes to the MSLP forecast are obtained in the ALL experiments, the smallest actual MSLP changes occur in the DRY experiments. Since the predicted MSLP changes are proportional to the amplitudes of the IC metric [cf. Eq. (6)], this result implies that the sensitivity of ALL and MOIST experiments is such that the larger (smaller) the perturbation to the initial lower-tropospheric moisture, the stronger (weaker) Katia is at 72 h. However, the actual MSLP changes of the DRY experiments are less than half the predicted MSLP changes, which indicates that perturbing only the kinematic field has comparatively little impact on the intensity evolution of Katia. Additional experiments were done with different IC metrics (e.g., 500-hPa MSE and 850-hPa circulation), but the greatest impact on the MSLP of Katia was attained with the 950-hPa MSE and when water mixing ratios were perturbed. These results further support the relevance of lower-tropospheric moisture to the right of the shear vector in the intensification of Katia.

9. Conclusions

This study investigated the intensity changes of Katia (2011) with a 96-member ensemble forecast from the AHW Model. The intensity forecasts, initialized when Katia was affected by moderate easterly shear, were characterized by large uncertainty. Composites of two groups of 12 members (designated as strong and weak members) from opposite ends of the forecast spectrum were compared to investigate the underlying mechanisms behind the different intensity evolutions in an environment characterized by moderate shear.

Comparing first the environmental conditions revealed that, even though the initially moderate shear magnitude decreased for all members from 0 to 18 h, the shear in strong members was significantly weaker than in weak members from 18 to 36 h. Strong members were also characterized by significantly greater precipitable water vapor from 0 to 84 h. Other environmental diagnoses were considered, but no consistent and statistical significance was found.

Further diagnosis of the moisture field was performed on a storm-centered framework to better understand where and when water vapor differences affected the evolution of Katia. The largest differences in moisture were found in the lower troposphere during the first
24 h, when strong members had more water vapor north, or right of shear, of Katia. The location of the water vapor differences was important, because air parcels that originated within the area of largest differences became near-saturated as they entered the inner circulation of Katia and ascended into deep updrafts. The increased lower-tropospheric moisture in the boundary layer favored buoyant updrafts that increased in

Fig. 14. Height–time analysis of the area-averaged vorticity within 100-km radius of Katia. The ensemble-mean differences between the (a) ALL, (b) MOIST, and (c) DRY experiments and the control simulation (shading; every $0.5 \times 10^{-3} \text{s}^{-1}$). Black contours represent the ensemble mean from the control simulation (every $2.5 \times 10^{-3} \text{s}^{-1}$).
magnitude and number over time. As the forecasts evolved, strong members went from having more low-level moisture to more midlevel moisture. Water vapor budgets revealed that greater horizontal flux convergence in the boundary layer and vertical flux convergence above, as well as less horizontal water vapor flux divergence at midlevels, aided strong members in moistening through the midtroposphere and having deeper latent heat release from convection.

Intensity changes of strong and weak members were further investigated with a vorticity budget. Changes in the area-averaged vorticity of strong members during
12–24 h were characterized by greater vortex stretching in the lower troposphere and tilting of the horizontal vorticity in the midtroposphere. As the midtroposphere of strong members became more moist between 36 and 48 h, deeper vortex stretching and increased tilting added up together to increase the circulation at all levels. This process of intensification is somewhat similar to the observational study of Dolling and Barnes (2012), which found that increased buoyancy and vortex stretching aided intensification of a weak storm. In the case of Katia, cyclonic stretching was first limited to a shallow layer, which helped to sustain the low-level circulation until deeper cyclonic stretching took place.

The findings of this study suggest that uncertainty in initial conditions in the moisture field have a significant impact on the subsequent forecast uncertainty. This result implies that observations from lower- and midtropospheric moisture, especially downshear and right of the shear, could help reduce initial conditions and forecast uncertainty. This idea was tested with sensitivity experiments in which the initial conditions were adjusted in response to observations of the 950-hPa MSE. An inverse relationship between the 950-hPa MSE and TC intensity was found such that an observation of larger MSE north, or right of shear, of Katia resulted in a lower 72-h MSLP forecast. Furthermore, the impact on the MSLP forecast was nearly similar when only water variables or all state variables were adjusted, further supporting the relevance of moisture certainty for intensity forecasts of weak, sheared storms.

Last, the findings presented here are only based on the evaluation of the AHW forecasts of Katia. To verify the results for robustness and to compare the intensity evolution in a different shear profile, Part II will present similar analyses for the AHW forecasts of Ophelia. Future research should compare the results of this study with observations of sheared TCs to elucidate the validity of the intensification processes identified here.

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FIG. 16. Comparison between the predicted and actual ensemble-mean 72-h MSLP changes within the 96-member AHW ensemble. Values on the ordinate represent the difference between the sensitivity experiments and the control simulation, while values on the abscissa represent the prediction from Eq. (6). Each symbol represents an individual experiment with a prescribed perturbation amplitude to the IC metric. Black dots, blue squares, and brown diamonds represent the ALL, MOIST, and DRY experiments, respectively. A dashed black line is shown to reference a perfectly linear agreement between the predicted and actual MSLP forecast change.


