ABSRACT

A statistical analysis of tropical cyclone (TC) environmental wind profiles is conducted in order to better understand how vertical wind shear influences TC intensity change. The wind profiles are computed from global atmospheric reanalyses around the best track locations of 7554 TC cases in the Northern Hemisphere tropics. Mean wind profiles within each basin exhibit significant differences in the magnitude and direction of vertical wind shear. Comparisons between TC environments and randomly selected “non-TC” environments highlight the synoptic regimes that support TCs in each basin, which are often characterized by weaker deep-layer shear. Because weaker deep-layer shear may not be the only aspect of the environmental flow that makes a TC environment more favorable for TCs, two new parameters are developed to describe the height and depth of vertical shear. Distributions of these parameters indicate that, in both TC and non-TC environments, vertical shear most frequently occurs in shallow layers and in the upper troposphere. Linear correlations between each shear parameter and TC intensity change show that shallow, upper-level shear is slightly more favorable for TC intensification. But these relationships vary by basin and neither parameter independently explains more than 5% of the variance in TC intensity change between 12 and 120 h. As such, the shear height and depth parameters in this study do not appear to be viable predictors for statistical intensity prediction, though similar measures of midtropospheric vertical wind shear may be more important in particularly challenging intensity forecasts.

1. Introduction

Vertical wind shear in the environment of a tropical cyclone (TC) plays a central role in its intensity evolution and predictability (Gray 1968; Tuleya and Kurihara 1981; DeMaria 1996; Wang and Holland 1996; DeMaria and Kaplan 1999; Frank and Ritchie 1999, 2001; Wong and Chan 2004; Riemer et al. 2010; Zhang and Tao 2013). As the magnitude of vertical wind shear increases, the environment is widely considered to be less favorable for TC formation and intensification. The destructive effect of vertical wind shear owes to a variety of proposed mechanisms, which include the dilution of the upper-level warm core (Gray 1968; Frank and Ritchie 2001), ventilation of the midlevel vortex with dry environmental air (Simpson and Riehl 1958; Cram et al. 2007; Tang and Emanuel 2010), and enhanced flushing of low-entropy air into the frictional inflow layer (Tang and Emanuel 2010; Riemer et al. 2010, 2013; Molinari et al. 2013; Riemer and Laliberté 2015). It has become common practice to use the deep-layer shear, computed as the difference between winds at the 200- and 850-hPa standard pressure levels, to estimate the extent to which these mechanisms act to limit TC formation or intensification. But as routine analyses and satellite observations are increasingly capable of resolving the vertical structure of the environmental winds, we seek to understand whether aspects of the environmental flow beyond the deep-layer shear have a meaningful effect on the intensity of real TCs. To this end, we develop two new parameters for describing the vertical distribution of the horizontal winds in TC environments, and we quantify their statistical relationship to TC intensity change in a large sample of storms. We also use this large sample of storms to understand basic characteristics and variability of wind profiles and wind shear in TC environments.

Several studies have used the deep-layer vertical wind shear in the tropics to understand patterns and variability of TC formation, and to predict intensity change. Aiyyer and Thornicroft (2006) developed a climatology of deep-layer shear over the tropical Atlantic from
45 years of reanalysis data, shedding light on the spatiotemporal patterns of TC genesis in this region. They extended their analysis throughout the tropics, linking modes of wind shear variability to seasonal fluctuations in TC activity in different basins (Aiyer and Thorncroft 2011). Using profiling observations from four locations in the Caribbean, Dunion (2011) characterized the favorableness of three prevailing air masses for TCs based partly on their average deep-layer shear. The deep-layer shear is also a key component of TC genesis parameters (Emanuel and Nolan 2004; McGauley and Nolan 2011) and statistical intensity prediction models (DeMaria and Kaplan 1994, 1999; DeMaria et al. 2005; Knaff et al. 2005).

Fewer studies have focused on how aspects of the environmental flow other than the deep-layer vertical wind shear affect TC intensity. This is somewhat surprising, as the two-level wind difference between 200 and 850-hPa may occasionally mislead the forecaster about the destructive potential of the environmental flow (Elsberry and Jeffries 1996; Velden and Sears 2014). Evidence of this can be found in both modeling studies and statistical analyses over large samples of real storms. Zeng et al. (2010) used TC best track and reanalysis data from the Atlantic to statistically analyze how vertical shear measured in different layers of the TC environment influences lagged intensity change. They found that the 200–850-hPa vertical shear does not always explain the most variance in the lagged intensity change. Wang et al. (2015, hereafter W15) extended this analysis to a sample of west Pacific typhoons and found that vertical shear in the lower troposphere correlates more strongly with weakening typhoons during the active part of the typhoon season. Xu and Wang (2013) and Finocchio et al. (2016, hereafter F16) exposed idealized TC-like vortices to different wind profiles with the same amount of deep-layer shear. Changing only the vertical distribution of the shear elicited significant TC intensity and structural responses. In agreement with W15, F16 found that environments with shallow and low-level vertical wind shear were less favorable for the intensification of model TCs. F16 proposed a pathway for explaining why low-level shear was more unfavorable for TCs. According to this pathway, low-level shear tilts the TC vortex more effectively and facilitates a radially inward intrusion of low-θ_e air from the midlevel environment. This, in turn, enhances boundary layer flushing akin to that identified by Riemer et al. (2010) and Molinari et al. (2013), which frustrates the realignment and intensification of the TC.

Similar to W15, this study investigates statistical relationships between TC intensity change and vertical shear in different layers of the troposphere. One difference from W15 is that our sample of storms spans most of the Northern Hemisphere tropics. This allows us to compare statistics of TC environmental winds and vertical wind shear between basins (section 3) and assess differences between TC flow environments and the mean “non-TC” state in each basin (section 4). Another difference from W15 is that we develop two new parameters for quantifying the height and depth of vertical wind shear (section 5). After controlling for relationships between these new parameters and the deep-layer shear, we find that the shear height and depth parameters are actually weak statistical predictors of TC intensity change.

2. Data and methods

We use a combination of TC best track and global atmospheric reanalysis data to construct the environmental wind profiles used throughout this study. The TC location and intensity information used to select the sample of storms comes from the International Best Track Archive for Climate Stewardship (IBTrACS v03r07; Knapp et al. 2010). Specifically, we use the IBTrACS-WMO subset that only contains best track data furnished by World Meteorological Organization (WMO) Regional Specialized Meteorological Centers or Tropical Cyclone Warning Centers.

We apply a variety of geographical selection criteria in order to obtain our sample of TCs from the IBTrACS-WMO archive. First, we only consider TCs within the three major ocean basins of the Northern Hemisphere: the western North Pacific from 100°E to 180° (WP), the eastern North Pacific from 180° to the west coast of the American continent (EP), and the North Atlantic from the east coast of the American continent to 10°W (AL). Our analysis focuses on the tropics, from the equator to 30°N, where cold SST and extratropical transition are less likely to influence TC intensity and structure. We also exclude TC cases that are closer than 500 km to landmasses in order to further minimize the influence of factors other than the atmospheric environment on the storms in our sample.

We utilize the best track wind speed information in order to select cases within the domain of interest that have maximum sustained winds at or above tropical storm strength (34 kt, 18 m s⁻¹). Best track winds in the AL and EP represent a 1-min maximum sustained surface wind speed, while best track winds for WP TCs are a 10-min maximum wind. So that the wind speed criteria are applied consistently across basins, we convert from 10- to 1-min winds by dividing WP best track wind speeds by 0.88 (Harper et al. 2008; Schreck et al. 2014). Even after converting to 1-min winds, IBTrACS-WMO wind speeds for WP TCs may still be somewhat weaker than those in the Joint Typhoon Warning Center best
track (Schreck et al. 2014), but we do not expect this to significantly affect case selection or the ensuing analysis. Finally, we only consider best track records at 0000 and 1200 UTC, from 1 May to 31 October, during the years 1979–2014 for which routine satellite observations are available. Figure 1 depicts the locations of the 7554 TC cases that satisfy these criteria, among which 3764 occur in the WP (49.8%), 2393 occur in the EP (31.7%), and 1397 occur in the AL (18.5%).

Global atmospheric reanalysis is used to characterize the environmental flow around each TC in the sample. In this study, we use the Japanese 55-year Reanalysis (JRA-55) produced by the Japan Meteorological Agency’s global atmospheric model and 4DVAR data assimilation system (Kobayashi et al. 2015). The data are horizontally interpolated from the native TL319 reduced Gaussian grid to a 1.25° latitude–longitude grid, and vertically interpolated from 60 native levels to 37 isobaric levels. JRA-55 has been shown to improve upon the representation of TCs in comparable reanalyses. For instance, Kobayashi et al. (2015) found TC detection rates in JRA-55 to be consistently higher than in the European Centre for Medium-Range Weather Forecasts (ECMWF) interim reanalysis, sometimes by as much as 30% prior to the 1990s. Moreover, in a comparison of six global reanalyses, Murakami (2014) found that JRA-55 produces more storms of hurricane intensity and more accurately represents the positions of TCs.

Although an adequate representation of TCs in reanalysis is desirable, the flow associated with a TC vortex can locally change the vertical wind shear around the storm. Because we are primarily interested in the vertical wind shear associated with the surrounding environment, we remove the vortex wind field for each case using the technique of Kurihara et al. (1993, hereafter KBR93). The vortex removal technique replaces the flow within a cylindrical region encompassing the TC by optimally interpolating the flow along the cylinder edge inward. Originally, KBR93 defined the cylinder radius using a set of criteria involving the radial profile of axisymmetric tangential winds at the top of the boundary layer ($V_d$). We use an adjusted set of rules, which defines the filter radius $r_f$ as the innermost radius (moving outward from 1.5 times the radius of maximum wind) at which 1) $V_d < 3 \text{ m s}^{-1}$ and $dV_d/dr > -2 \times 10^{-6} \text{s}^{-1}$ for four consecutive radial rings, 2) $V_d < 1.5 \text{ m s}^{-1}$, or 3) the distance to the storm center in 1.25° reanalysis grid points reaches the smaller of 8 and 1/10 of the best track maximum wind speed in knots. This last rule, in which the filter radius is made to be a linear function of best track maximum wind speed, is intended to preserve the environmental flow closer to the center of weaker storms. Finally, we do not inflate $r_f$ determined from these adjusted criteria, as was done in KBR93, because the uninflated $r_f$ sufficiently captures the vortex. KBR93 defined the center of the cylindrical filter domain using a wind speed centroid. In this study, we assume the best track position is approximately the same as the reanalysis storm center (and thus, the filter domain center) position. This is justified by Harada et al. (2016), who found that root-mean-squared differences between the best track and JRA-55 TC center positions were generally smaller than the 1.25° grid spacing of the reanalysis.

Optimal interpolation within the resulting cylindrical region produces what is referred to as the nonhurricane wind field because it removes only the part of the wind field associated with the TC vortex. Figure 2 depicts an example of the total and nonhurricane wind fields at 850 and 200 hPa for Atlantic Hurricane Isabel (2003). Filtering removes Isabel’s strong cyclonic circulation at 850 hPa and the divergent outflow at 200 hPa, while retaining the surrounding disturbances that contribute to the mean vertical wind shear in Isabel’s environment.

The mean environmental wind profile corresponding to each valid TC case in the sample is computed by spatially averaging the nonhurricane part of the zonal and meridional winds at each vertical level. Spatial averaging takes place in an annulus extending 200–800 km from the best track position—the same annulus that has been used to compute many environmental fields in statistical intensity prediction models (DeMaria et al. 2005; Knaff et al. 2005).

Our extensive use of reanalysis necessitates some practical consideration of the errors associated with these data. In addition to the reanalysis being too coarse to simulate...
finer-scale TC responses to vertical wind shear, JRA-55 wind fields have inherent errors and biases. Using 6-hourly rawindsondes released from 22 sites in southeastern China during the warm season, Chen et al. (2014) found RMS errors in both the $u$ and $v$ winds of about 2 m s$^{-1}$ below 400 hPa, which increased to 2.5 m s$^{-1}$ at 200 hPa. The magnitudes of northerly and easterly wind biases in the lower troposphere were never greater than 0.5 m s$^{-1}$ and decreased to near 0 m s$^{-1}$ at 200 hPa. It should be noted that these error statistics were not obtained from an independent validation because the validating rawindsondes were assimilated in JRA-55. Furthermore, the high spatiotemporal frequency of wind profiling observations in their study would not typically occur in TC environments. It is also unclear how the error statistics in southeastern China compare to JRA-55 global error statistics. Therefore, TC environmental winds diagnosed from JRA-55 are likely to have somewhat larger RSME and biases than the values reported in Chen et al. (2014).

3. Interbasin comparison

a. Mean wind profiles

The mean zonal and meridional wind profiles for valid TCs in each basin are shown in Fig. 3. So that the comparison of mean wind profiles is independent of the different sample sizes, we compute 200 additional mean wind profiles in each basin from sets of 500 randomly selected cases. The shaded regions in Fig. 3 encapsulate 98% of these random mean profiles that are closest to the mean over all profiles in each basin, and are henceforth referred to as the 98th percentile regions. We only consider two basin mean profiles to be statistically different from one another if their 98th percentile regions do not overlap. The WP and EP mean zonal wind profiles are statistically similar at 1000 hPa and above 150 hPa. The AL mean zonal wind profile is statistically similar to the EP from 400 to 650 hPa, and to the WP from 350 to 200 hPa. Otherwise, there are significant differences between the mean zonal wind profiles that translate to differences in the mean deep-layer shear in each basin. Mean westerly vertical shear throughout the troposphere is more characteristic of the AL; the magnitude of deep-layer shear in the mean zonal wind profile is 6.2 m s$^{-1}$ in the AL versus 1 m s$^{-1}$ in the EP and 2.2 m s$^{-1}$ in the WP. The EP exhibits a shallow layer of weak westerly shear between 600 and 250 hPa, with easterly shear below and above this layer. The WP has almost no mean vertical shear of the zonal winds through the
middle troposphere, with easterly shear above 200 hPa and westerly shear below 700 hPa.

The mean meridional wind profiles are also significantly different from one another in all three basins, but generally exhibit weaker wind speeds at all levels and less vertical wind shear. For example, the mean meridional winds are no greater than 6 m s$^{-1}$ at any level in the EP and are slightly southerly but always 2 m s$^{-1}$ in the AL. The mean meridional wind profile in the WP, however, has a northerly deep-layer shear of 3.2 m s$^{-1}$.

Considering the profile of $y$-wind biases in JRA-55 exhibits weak southerly shear through the middle levels (Chen et al. 2014), this mean northerly shear is not an artifact of the reanalysis data, but instead reflects the mean conditions where TCs tend to occur in the WP: southerly low-level flow in the monsoon trough underlying northerly upper-level flow on the eastern periphery of the South Asian high (Wang and Wu 2016).

b. Wind profile distributions

Distributions of TC environmental wind profiles reveal the variability in the environmental flows to which TCs are exposed in the Northern Hemisphere. Figure 4 depicts contour frequency by altitude diagrams (CFADs) constructed from the zonal and meridional wind profiles around TCs in each basin. The mean wind profiles from Fig. 3 appear as solid black lines in each panel. Similar to the means in all three basins, the modes of the zonal wind distributions in the lower troposphere are easterly and generally weaker than 5 m s$^{-1}$. Westerly surface winds are particularly rare in the AL and are almost never stronger than 5 m s$^{-1}$ in the EP and WP. Despite smaller spread in meridional winds at all levels, both zonal and meridional wind distributions in each basin are broadest near 200 hPa.

We can gain a more quantitative sense for differences in the wind distributions among basins by computing the standard deviation (i.e., “spread”) of the zonal and meridional winds at each vertical level (Fig. 5). The shaded region around each basin’s wind spread profile in Fig. 5 is computed in the same way as the shaded region around the mean profiles in Fig. 3. Zonal and meridional wind spread increases with height in each basin such that the spread at 200 hPa is anywhere from 2 to 3 times larger than that at 850 hPa. The meridional wind spread remains fairly constant below 400 hPa, but increases sharply above this level. Differences in the meridional wind spread among basins are most significant above 200 hPa where the AL and WP exhibit larger spread than the EP. Zonal wind spread is consistently more than 25% larger than meridional wind spread above 850 hPa, and is more basin dependent. For example, zonal wind spread in the WP is more than 1 m s$^{-1}$ larger than in the EP at all levels above 500 hPa and increases almost linearly with height (in the depicted log-pressure coordinate). In contrast, zonal wind spread in the EP only begins increasing with height above 500 hPa. Despite these differences, the upper-level maximum in zonal and meridional wind spread in all three basins generally confirms previous studies that found the variability in deep-layer shear to be controlled by the upper-level flow (Gray 1968; Goldenberg and Shapiro 1996).

c. Deep-layer vertical wind shear distributions

Given the importance of deep-layer shear in predicting TC intensity change, we briefly compare distributions of
this bulk metric before exploring more detailed aspects of vertical wind shear in section 5. Solid lines in Fig. 6 show distributions of each component of deep-layer shear in TC environments. Comparing just the zonal component of deep-layer shear, it is apparent that the AL distribution is shifted to significantly larger westerly shear than the EP or WP distributions. More frequent westerly shear in the AL agrees with the larger westerly shear evident in this

FIG. 4. CFADs of (left) zonal and (right) meridional winds computed over cases within the (top) WP, (middle) EP, and (bottom) AL. Color shading denotes the frequency distribution (%) of winds at each vertical level. Solid black lines in each panel are the mean wind profiles and dashed black lines are ±1 standard deviation from the mean wind profile.
basin’s mean zonal wind profile (Fig. 3). Distributions of the meridional component of deep-layer shear are more peaked at small magnitudes, indicating that the zonal wind component is the larger contributor to deep-layer vertical wind shear in TC environments. In agreement with the mean meridional wind profile in Fig. 3, the distribution of the meridional component of deep-layer shear in the WP shows higher frequencies of northerly shear.

The relationship between deep-layer shear and TC intensity change is evident in how the distributions change when we consider only intensifying storms (dashed lines in Fig. 6). We define intensifying storms as cases in which the best track wind speed increases by at least 5 m s$^{-1}$ and the best track pressure decreases by at least 2 hPa in two consecutive 12-h intervals from 0 to 12 h and from 12 to 24 h. Table 1 lists the number of intensifying cases in each basin. In general, the distributions for intensifying storms are more peaked such that the largest shear magnitudes are less frequent. The AL zonal distribution actually shifts toward the weaker westerly shear, which suggests that the strong westerly shear is at least partly responsible for the large percentage of weakening or steady-state storms in this basin. Meridional shear distributions also become

![Fig. 5. Standard deviation of (left) zonal and (right) meridional winds and corresponding 98th percentile regions (shading) for TC environments in each basin.](image)

![Fig. 6. Probability distributions of (left) zonal and (right) meridional components of the deep-layer vertical wind shear in each basin. Solid lines are distributions for all TCs and dashed lines are distributions for only intensifying TCs.](image)
more peaked for intensifying storms, with a noticeable shift in the EP distributional peak from nearly zero to light northerly shear. It is unclear whether this shift has a physical explanation, or is simply due to computing distributions over a smaller sample of storms.

4. TC versus non-TC environments

To illustrate the effect of anchoring our environmental wind analysis to the locations of TCs, we compare the TC flow environments from section 3 to flow environments around randomly selected locations throughout the tropical basins considered in this study. Our method of obtaining the sample used to construct what we henceforth refer to as the non-TC wind profile distribution is based on an approach developed by Nolan and McGauley (2012). For each valid TC case, we randomly select the locations of 40 other valid TC cases that occur within the same basin, but at least 5° away from the current case and during different years. Following the method for computing the mean wind profile around TCs, but without removing a vortex, we compute the mean wind profile around each of these 40 points in the basin using the reanalysis data at the time of the current TC case. The resulting wind profiles correspond to realistic TC locations that are sufficiently far from the current TC case, earning their designation as “non-TC” environments. Repeating this procedure for all of the valid TC cases produces a non-TC sample that is 40 times larger than the TC sample. Aside from the non-TC wind distributions being broader at all vertical levels, differences between TC and non-TC wind profile distributions are difficult to detect visually using CFADs (not shown). Nevertheless, a Kolmogorov–Smirnov test indicates with 98% confidence that the TC and non-TC wind distributions are statistically different from one another at all levels in each basin, except from 250 to 350 hPa for AL zonal winds.

For assessing the significance of differences between mean TC and non-TC wind profiles in each basin, we again rely on 200 mean wind profiles computed over different subsets of randomly selected non-TC cases in each basin. Each subset contains the same number of non-TC cases as there are valid TC cases in each basin so that the comparison between TC and non-TC wind profiles is independent of sample size. Figure 7 depicts the mean non-TC wind profiles (dashed lines) and their 98th percentile regions alongside the same mean TC wind profiles (solid lines) and 98th percentile regions from Fig. 3. Differences in the mean zonal winds at 200 hPa in the AL are not significant according to Fig. 7. However, the 850-hPa easterlies are significantly weaker in AL TC environments such that the 0.8 m s⁻¹ reduction in the westerly deep-layer shear relative to non-TC environments is statistically significant. TC environments in the EP also have weaker easterly flow at 850 hPa, which, combined with the light easterly flow at 200 hPa, results in a 3.5 m s⁻¹ reduction in the westerly deep-layer shear relative to non-TC environments. The mean winds through the lower and middle troposphere of WP TC environments are significantly less easterly and more southerly than in non-TC environments, but the deep-layer shear magnitude is only 0.4 m s⁻¹ weaker. Despite the small reductions in the mean deep-layer shear in the WP and AL, distributions of deep-layer shear for non-TC samples (not shown) depict higher probabilities of deep-layer shear exceeding 20 m s⁻¹ in all three basins.

Diagnosing differences in deep-layer shear through the comparison of mean wind profiles does not account for how the spatial distribution of shear changes in TC environments. Using the AL as an example, we illustrate such distributional changes by averaging the environmental deep-layer shear referenced to TC and non-TC positions within 1° × 1° latitude–longitude bins. In non-TC environments (Fig. 8a), there is a tongue of deep-layer shear exceeding 12 m s⁻¹ in the central and eastern AL with a sharp transition to weaker shear south of 15°N. This pattern matches the climatological deep-layer shear pattern in Fig. 3 of Aiyyer and Thorncroft (2006) and in Fig. 3 of Goldenberg and Shapiro (1996). Shear magnitudes markedly decrease within the main development region when the shear is computed in TC environments (Fig. 8b). And if we only consider the 308 intensifying storms (Fig. 8c), then the reduction in shear magnitude is even more pronounced and the distribution of cases becomes more confined to the main development region. Such significant reductions in deep-layer shear relative to the mean non-TC state in the AL speak to the

<p>| TABLE 1. Number of TC cases and intensifying TC cases in each basin, and the percentage of TC cases in each basin that are intensifying. |
|-----------------------------------|-----|-----|-----|-----|</p>
<table>
<thead>
<tr>
<th></th>
<th>WP</th>
<th>EP</th>
<th>AL</th>
<th>Total</th>
</tr>
</thead>
<tbody>
<tr>
<td>TCs</td>
<td>3764</td>
<td>2393</td>
<td>1397</td>
<td>7554</td>
</tr>
<tr>
<td>Intensifying TCs</td>
<td>1120</td>
<td>474</td>
<td>308</td>
<td>1902</td>
</tr>
<tr>
<td>Percent of TCs intensifying</td>
<td>29.8%</td>
<td>19.8%</td>
<td>22.0%</td>
<td>25.2%</td>
</tr>
</tbody>
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infrequency of favorable TC conditions in this basin, as pointed out by Aiyyer and Thorncroft (2006).

5. Vertical distribution of vertical wind shear

The idealized simulations of F16 showed that, in addition to the deep-layer shear, the vertical distribution of horizontal winds can strongly influence TC intensity change. Our goal in this section is to determine whether such sensitivity to the vertical distribution of vertical wind shear can be detected in our sample of real storms.

a. Shear height and depth parameters

The two parameters we use to describe the vertical distribution of shear utilize the local vertical wind shear at each level between 200 and 850 hPa. The zonal component of vertical shear at a pressure level \( p_i \), for example, is the centered difference between the zonal wind at the reanalysis level just above \( (p_{i+1}) \) and below \( (p_{i-1}) \) that level: 
\[
\frac{\partial u}{\partial p}|_{p_i} \approx \frac{u(p_{i+1}) - u(p_{i-1})}{2\Delta p},
\]
where \( \Delta p = 50 \) hPa. The local shear magnitude at pressure level \( p_i \) is then 
\[
\left[ (\frac{\partial u}{\partial p}|_{p_i})^2 + (\frac{\partial v}{\partial p}|_{p_i})^2 \right]^{1/2}.
\]

The shear depth parameter \( d \) is assigned as the depth (hPa) of the deepest layer between 200 and 850 hPa in which the local shear magnitude continuously exceeds a threshold. Global thresholds can prevent wind profiles with particularly weak shear from being assigned a valid shear depth, so we set the threshold to be 66% of the average local shear magnitude between 200 and 850 hPa in each profile. The 66% threshold is large enough to discern cases with deep layers of significant shear from the many cases with light but nontrivial shear at all levels. At the same time, the 66% threshold is small enough to be exceeded on at least two consecutive vertical levels of every profile in the sample, guaranteeing that each case is assigned a valid shear depth.

![Figure 7](image1.png)

FIG. 7. As in Fig. 3, but with the addition of mean wind profiles over the sample of non-TC cases in each basin (dashed lines).

![Figure 8](image2.png)

FIG. 8. Deep-layer vertical wind shear averaged within \( 1^\circ \times 1^\circ \) bins in the AL for (a) non-TC, (b) TC, and (c) intensifying TC cases.
The left panel of Fig. 9 depicts the distributions of $d$ for TC (solid) and non-TC (dashed) environments. Both TC and non-TC distributions peak around 200 hPa, indicating a higher frequency of relatively shallow shear. The slight shift of the non-TC distribution to larger $d$ is due to non-TC environments having stronger shear, which we will show in section 5b correlates with deeper shear. The right panel of Fig. 9 depicts composite hodographs for each sample of cases. Each hodograph depicts the winds between 850 and 200 hPa relative to the 850-hPa wind. Points corresponding to the 500-, 350-, and 200-hPa winds are labeled.

The second parameter $h$ describes the height of vertical wind shear. The first step in computing $h$ is obtaining a cumulative shear distribution (CSD) for each wind profile. We compute a wind profile’s CSD ($CSD_{prof}$) by dividing the accumulated local shear at each level from 200 hPa to a pressure level $p$ by the sum of shear magnitudes over all levels between 200 and 850 hPa. Note that $CSD_{prof}$ does not take into account the shear direction. The gray lines in Fig. 10a depict CSDs for all TC environments in the sample. More concave (convex) CSDs correspond to wind profiles with vertical shear primarily in the upper (lower) levels. The dashed black line in Fig. 10a denotes a CSD for a hypothetical wind profile in which the total shear between 200 and 850 hPa is assumed to be equally partitioned among pressure levels. This linear CSD is expressed as

$$CSD_{lin}(p) = \frac{p - 200 \text{ hPa}}{850 - 200 \text{ hPa}}.$$  \hspace{1cm} (1)

Adding the differences between $CSD_{lin}$ and $CSD_{prof}$ at each level from 200 to 850 hPa yields the dimensionless shear height parameter:

$$h = \sum_{p=200 \text{ hPa}}^{850 \text{ hPa}} \frac{\text{CSD}_{prof}(p) - CSD_{lin}(p)}{\text{CSD}_{prof}(p)}.$$  \hspace{1cm} (2)

This parameter quantifies how each wind profile’s cumulative shear distribution deviates from the evenly distributed shear case. Larger positive (negative) values of $h$ correspond to wind profiles with shear confined to the upper (lower) levels. Figure 10b shows the distributions of $h$ for the TC (solid) and non-TC (dashed) samples. Both distributions are remarkably similar, with a peak around $h = 2$ indicating a preference for upper-level shear that is not unique to TC environments. The slight shift of the non-TC distribution to larger $h$ is, again, due to non-TC environments having stronger shear, which also weakly correlates with larger values of $h$ (section 5b).

To better illustrate the correspondence between $h$ and the height of shear in TC environments, we select three
wind profiles with $h > 4$ (blue), $h = 0$ (purple), and $h < 0$ (red) and plot their corresponding hodographs (Fig. 10c).
The larger spacing between the four dots corresponding to the 350–200-hPa winds on the blue hodograph illustrates how most of the shear indeed occurs in the upper levels for profiles with larger values of $h$. The percentage of total shear above 500 hPa also shows the close correspondence between $h$ and the height of vertical wind shear; the profile with $h > 4$ (blue) has 88% of its total shear above 500 hPa, while the profiles with $h = 0$ (purple) and $h < 0$ (red) have 46% and 35%, respectively, of their total shear above 500 hPa.

From the hodographs in Fig. 9, we notice a relationship between the shear height and depth parameters; as the shear depth parameter decreases, the composite hodographs in Fig. 9 have increasingly upper-level shear. Linear correlations between $h$ and $d$ computed over the combined sample of storms indeed show a weak anticorrelation between the two parameters. The anticorrelation is larger in the WP, where $d$ explains 8% of the variance in $h$, than in the AL and EP, where $d$ explains 1.6% and 0.6% of the variance in $h$, respectively. The overall weakness of the relationship between the shear height and depth parameters, however, suggests that these two parameters each describe sufficiently unique structural characteristics of vertical wind shear so as to be treated separately in this study.

b. Relationship with deep-layer vertical wind shear

The remainder of this study aims to quantify the skill that the two new shear parameters can add to TC intensity forecasts relative to the widely used deep-layer shear (DSHR). This section establishes the relationship between DSHR and each variable by comparing geographic distributions and computing linear correlations between DSHR and the new shear parameters. These correlations measure the amount of information the shear parameters describe about the environmental flow that is potentially redundant with the deep-layer shear.

Figure 11 shows geographic distributions of $d$, $h$, and DSHR. There are rather remarkable differences in the shear height parameter by basin (Fig. 11b). Upper-level shear is more prevalent in the WP, where the mean value of $h$ (2.3) is 18% larger than in the AL and 32% larger than in the EP. Although differences in shear depth between basins from Fig. 11a are not as obvious, a relationship with DSHR is evident when we compare Figs. 11a and 11c. The deepest shear tends to occur in the same regions as the strongest deep-layer shear magnitude: in the central Atlantic outside of the main development region and in the central Pacific. Furthermore, the largest mean values of DSHR (9.0 m s$^{-1}$) and $d$ (313.8 hPa) both occur in the AL, while the smallest
mean values of $\text{DSHR}$ (6.5 m s$^{-1}$) and $d$ (289.5 hPa) both occur in the EP.

We quantify statistical relationships between DSHR and the new shear parameters using linear correlations. Rather than computing correlations between variables at an instant in time, our approach throughout this study is to compute correlations between variables averaged along the best track within lag time intervals from 12 to 120 h. This averaging accounts for time variations in the TC environment (DeMaria and Kaplan 1994). The resulting time series of correlation coefficients at each lag also shows how statistical relationships between variables evolve. Valid TC cases must satisfy the case selection criteria described in section 2, with the only difference being that storms at lag times are allowed to come as close as 200 km to land. The number of valid cases at each lag time appears in Table 2.

We assess the statistical significance of correlations using a two-tailed Student's $t$ test and a bootstrap test. The bootstrap test assigns as a confidence level the percent of regression lines computed for 100 randomly selected halves of the sample whose slopes have the same sign as the regression line for the full sample. We require both tests to indicate significance at the 98% confidence level.

The left panels in Fig. 12 scatter shear height and depth against DSHR. Each scattered variable is averaged within a 48-h lag interval. Within this interval, shear height exhibits a weak ($<$0.3) correlation with DSHR in the AL, EP, and combined sample, such that upper-level shear tends to be associated with stronger shear. In the WP, there is a weak anticorrelation, which is not statistically significant. The right panels of Fig. 12 depict time series of the correlation coefficients in each basin and in the combined sample (black line). The weak correlation between shear height and DSHR at 48-h lag is similar to correlations at other times. Unlike shear height, shear depth exhibits a moderate ($\geq 0.3$) correlation with DSHR after 24 h. This statistical relationship combined with the geographical congruence of DSHR and $d$ in Figs. 11a and 11c suggests that deeper shear tends to be associated with stronger shear.

c. Partial correlation analysis

We have shown that the shear height and depth parameters are related to the deep-layer vertical wind shear and, to a lesser extent, one another. These relationships make it difficult to ascribe changes in TC intensity exclusively to the height or depth of vertical wind shear, so we attempt to statistically isolate the intensity response to each parameter using partial correlations (Sippel and

**TABLE 2. Number of valid TC cases at each lag time (h).**

<table>
<thead>
<tr>
<th>Basin</th>
<th>12</th>
<th>24</th>
<th>36</th>
<th>48</th>
<th>72</th>
<th>96</th>
<th>120</th>
</tr>
</thead>
<tbody>
<tr>
<td>All</td>
<td>6819</td>
<td>6451</td>
<td>5623</td>
<td>5160</td>
<td>3978</td>
<td>2946</td>
<td>2091</td>
</tr>
<tr>
<td>AL</td>
<td>1267</td>
<td>1127</td>
<td>998</td>
<td>881</td>
<td>664</td>
<td>487</td>
<td>347</td>
</tr>
<tr>
<td>EP</td>
<td>2365</td>
<td>2288</td>
<td>2162</td>
<td>2002</td>
<td>1628</td>
<td>1262</td>
<td>936</td>
</tr>
<tr>
<td>WP</td>
<td>3189</td>
<td>3036</td>
<td>2463</td>
<td>2277</td>
<td>1686</td>
<td>1197</td>
<td>808</td>
</tr>
</tbody>
</table>
Zhang 2010; Sippel et al. 2011; Munsell et al. 2013). Partial correlations express the amount of variance in the predictand $y$ that is uniquely explained by a predictor $x$, after controlling for relationships between $x$ and other environmental variables. First-order partial correlation controls for a single variable $z$, and is expressed as

$$r_{y(xz)} = \frac{r_{yx} - r_{xz}r_{yz}}{\sqrt{1 - r_{xz}^2}},$$  

(3)

where $r$ denotes the correlation coefficient. The second-order partial correlation controls for relationships between $x$ and two variables $z_1$ and $z_2$, and is written in terms of first-order partial correlations as

$$r_{y(xz_1z_2)} = \frac{r_{y(xz_2)} - r_{y(z_1z_2)}r_{z_2(z_1)}}{\sqrt{1 - r_{z_2(z_1)}^2}}.$$  

(4)

In the present study, the predictand $y$ is the change in minimum sea level pressure $-\Delta\text{MSLP}$ along the best track within lag time intervals out to 120 h. We use $-\Delta\text{MSLP}$ so that positive values denote intensifying storms. The first control variable $z_1$ is DSHR and the...
second control variable \( z_2 \) is the new shear parameter that is not being used as the predictor. Thus, the second-order partial correlations measure the amount of variance in TC intensity change explained by shear height (depth) after controlling for changes in the deep-layer shear and shear depth (height). We assess statistical significance of partial correlations using a two-tailed Student’s \( t \) test in which \( t \) values are computed as

\[
t = r_{y(z_2, \ldots, z_k)} \sqrt{\frac{n - 2 - k}{1 - r^2_{y(z_2, \ldots, z_k)}}}
\]

Here, \( n \) is the number of cases at each lag time and \( k \) is the number of control variables, which equals two. We require significance at the 98% confidence level.

Figure 13 shows the time series of full correlations (solid lines) and second-order partial correlations between \( d \) and \(-\Delta \text{MSLP}\) that control for DSHR and \( h \) (dashed lines). Negative full correlations in all three basins indicate that deeper shear is less conducive for TC intensification, contrary to F16. The magnitude of this anticorrelation universally decreases when we control for DSHR and \( h \). The largest decreases (\( >40\% \)) occur in the AL, where deep-layer shear has the strongest influence on intensity change (not shown) and where there is a moderate correlation between DSHR and \( d \) (Fig. 12d). Even though shear depth alone explains less than 5% of the variance in intensity change at any time, the statistically significant partial correlations still suggest that deeper shear is slightly less conducive for intensification.

Full correlations between shear height and intensity change (Fig. 14, solid lines) are statistically significant at most lag times in the EP, WP, and combined sample. It is noteworthy that upper-level shear is associated with weakening TCs in the EP, but strengthening TCs in the combined sample and WP. We can invoke the geographical distributions of \( h \) (Fig. 11b) to explain the opposing signs of the full correlations. In the EP, the highest values of \( h \) occur north of 15°N, where the SST tends to become less supportive of TC intensification. Likewise, the highest values of \( h \) in the AL occur outside of the main development region, where the deep-layer shear is stronger. The coincidence of these hostile environmental conditions with upper-level shear, which should be more supportive of TC intensification according to F16, may explain the anticorrelation between \( h \) and intensity change that we observe in the AL and EP.

After controlling for its relationship with both DSHR and \( d, h \) exhibits a slightly weaker relationship with intensity change in all three basins. In the EP, the partial correlation increases by 10%–15% relative to the full correlation, denoting a weaker negative relationship with intensity change. Partial correlations decrease slightly in the WP, but remain similar to full correlations in the combined sample. These smaller changes suggest that the intensity response to shear height is not as closely tied to concomitant changes in the control variables as the intensity response to shear depth. Nevertheless, shear height alone only explains up to 2% of the variance in TC intensity change. So similar to shear depth, the shear height parameter is not a useful predictor of TC intensity change for the sample of storms in this study.

6. Discussion and conclusions

We used a large sample of TC cases across the Northern Hemisphere tropics to statistically analyze TC flow environments. Although we restricted our analysis to include only the active months in the Northern Hemisphere tropics, the sample contained enough diversity to clearly identify differences in the means and distributions of TC environmental wind profiles between basins. This large and diverse set of wind profiles provides realistic constraints for designing idealized wind profiles in future modeling studies.

We found significant differences in mean wind profiles and deep-layer shear distributions among basins and between TC and non-TC environments. On average, AL TCs are exposed to stronger westerly deep-layer shear than TCs in other basins. In the EP, and to a lesser extent in the AL and WP, the mean deep-layer shear in TC
environments is weaker than the mean shear when TCs are absent. This result demonstrates how climatological shear calculations that are not anchored to TC environments can overestimate the vertical wind shear in which TCs form and intensify (Nolan and McGauley 2012). Some of the diminished mean vertical wind shear in TC environments is due to weaker easterly trade winds, which may reflect the global preference for TCs during the convectively active phase of the Madden–Julian oscillation (Klotzbach 2014). TC environments in the WP, for which the mean vertical wind shear is not significantly weaker than in non-TC environments, tend to exhibit other differences that indicate synoptic regimes known to be favorable for TCs. For example, the stronger southerlies and weaker easterlies in the lower and middle levels of WP TC environments likely reflect the high frequency of typhoon development within the monsoon trough (Ritchie and Holland 1999; Molinari and Vollaro 2013) and perhaps to a lesser extent, within the weaker zonal background flows that support Rossby wave train formation in the wake of existing typhoons (Fu et al. 2007).

The spread in TC environmental wind profile distributions universally increases with altitude, but the explanations for this vary by basin. Although we restrict our sample to the tropics, transient midlatitude troughs can still contribute to larger upper-level wind spread, particularly in basins like the AL and WP that support TCs at higher latitudes. In the AL and EP, the enhanced upper-level wind spread could also be attributed to interannual variability associated with El Niño–Southern Oscillation and decadal variability associated with Sahel rainfall and the Atlantic meridional and Pacific decadal oscillations (Goldenberg and Shapiro 1996; Aiyyer and Thorncroft 2011). The larger wind spread in the upper levels of WP TCs likely reflects interannual variability in the position of the tropical upper-tropospheric trough (Wang and Wu 2016), and the higher frequency of binary TC interaction within this basin (Dong and Neumann 1983). Both of these factors may also explain the prevalence of upper-level wind shear in the WP (Fig. 11b).

Finally, we suspect the relatively smaller wind spread at all levels in the EP is due to the narrow corridor of favorable SST in this basin limiting the geographic coverage of TC cases.

Having larger wind spread in the upper levels implies that the variability of deep-layer shear in TC environments—a key predictor of intensity change—is controlled primarily by variability in the 200-hPa winds. This is not a particularly novel result, as several previous studies have tied the variability in the deep-layer shear to the upper-level flow (Gray 1968; Fitzpatrick et al. 1995; Goldenberg and Shapiro 1996; Aiyyer and Thorncroft 2006). Nevertheless, this result emphasizes the importance of upper-level wind measurements for estimating the deep-layer shear. Satellite-derived atmospheric motion vectors (AMVs) are often the only upper-tropospheric wind observations available in TC environments. But larger AMV height assignment errors are apt to occur when tracking thin cirrus tracers in the presence of strong vertical wind shear, which degrades the quality of upper-level AMVs (Velden and Bedka 2009; Sears and Velden 2012). These errors would translate directly into errors in deep-layer shear estimates and the TC intensity forecasts that rely on such estimates.

Motivated by recent evidence that the vertical distribution of vertical wind shear can significantly influence TC intensity, we devised two new parameters describing the depth and height of vertical shear. These parameters are similar to $\delta p$ and $p_{\text{max}}$ examined in the idealized simulations of F16, but are better suited for describing complex vertical shear profiles in real TC environments. Both parameters have narrow, Gaussian-like distributions indicating a propensity for shallow layers of upper-tropospheric wind shear in both TC and non-TC environments. Therefore, the idealized environments in F16 with the deepest and lowest-level vertical wind shear are the least likely to occur in nature.

We used linear correlation analysis to quantify statistical relationships between 12 and 120-h lagged intensity change and the shear height and depth parameters. The shear depth parameter was weakly anticorrelated with intensity change such that deeper shear was less conducive
for intensification. Partial correlations that control for the relationship between shear depth and magnitude weakened relative to the full correlations, but still remained negative. This result contradicts the idealized simulations of F16. We suspect that covariances between shear depth and other environmental predictors of TC intensity change not examined herein may explain this contradictory result. Furthermore, the apparent favorableness of more deeply distributed shear in F16 may have been a consequence of their idealized wind profiles not being representative of what real TCs experience. For the shear height parameter, linear correlations were weakly positive in the WP and combined sample of storms, weakly negative in the EP, and statistically insignificant in the AL. Controlling for the relationship between shear height and deep-layer shear slightly weakened the correlation with intensity change in all basins. In agreement with F16, upper-level shear was slightly more favorable for intensification in the WP and combined sample. The opposing relationship in the EP is likely due to upper-level shear occurring over cooler SST in this basin.

What is quite clear from Figs. 13 and 14 is that the new shear parameters are only weak predictors of TC intensity change. Based on the relatively narrow distributions of each shear parameter, one possible explanation is that their variability is too small to have an appreciable effect on real TCs. It is also possible that the shear parameters in this study do not adequately describe the vertical distribution of shear, which motivates further investigation into new parameters. Finally, the coarse resolution of our reanalysis data may contribute to the weak statistical relationships by only crudely capturing the processes by which shear height and depth modulate intensity change. To verify this, a similar statistical analysis should be repeated using a large set of real TCs simulated at high enough resolution to capture such physical processes as those identified by F16.

Notwithstanding the limitations of the shear parameters and reanalysis, our results suggest that including the height and depth of vertical wind shear in statistical intensity prediction models would not considerably improve forecasts. However, the height and depth of vertical wind shear may become more influential in scenarios known to have lower predictability, such as TC–trough interactions (Leroux et al. 2016) or TCs exposed to moderate wind shear in an otherwise favorable environment (Zhang and Tao 2013; Rios-Berrios et al. 2016). The prevalence of warm SST and elevated moisture in the west Pacific suggests that the latter scenario is common there. This could explain why W15 found clearer statistical relationships between intensity change and the layer in which vertical wind shear is measured in the west Pacific than did Zeng et al. (2010) in the Atlantic. Weak disturbances such as tropical waves and depressions, which are more exposed to the surrounding environment but were excluded from our sample, may also be more sensitive to subtle changes in the shape of a sheared environmental wind profile (Riemer and Montgomery 2011). So despite the weak statistical relationships for the diverse sample of storms in the present study, the utility of midlevel wind observations capable of resolving the structure of environmental vertical wind shear should not be discounted. We plan to conduct data assimilation experiments in different types of flow environments in order to quantify the effect of such observations, from both current and future platforms, on TC intensity forecasts.

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