Relation between Tropical Easterly Waves, Convection, and Tropical Cyclogenesis: A Lagrangian Perspective

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ABSTRACT

In this study, a wave-following Lagrangian framework was used to examine the evolution of tropical easterly wave structure, circulation, and convection in the days leading up to and including tropical cyclogenesis in the Atlantic and east Pacific basins. After easterly waves were separated into northerly, southerly, trough, and ridge phases using the National Centers for Environmental Prediction–National Center for Atmospheric Research reanalysis 700-hPa meridional wind, waves that developed a tropical cyclone [developing waves (DWs)] and waves that never developed a cyclone [nondeveloping waves (NDWs)] were identified. Day zero (D0) was defined as the day on which a tropical depression was identified for DWs or the day the waves achieved maximum 850-hPa vorticity for NDWs. Both waves types were then traced from five days prior to D0 (D−5) through one day after D0. Results suggest that as genesis is approached for DWs, the coverage by convection and cold cloudiness (e.g., fractional coverage by infrared brightness temperatures ≤240 K) increases, while convective intensity (e.g., lightning flash rate) decreases. Therefore, the coverage by convection appears to be more important than the intensity of convection for tropical cyclogenesis. In contrast, convective coverage and intensity both increase from D−5 to D0 for NDWs. Compared to NDWs, DWs are associated with significantly greater coverage by cold cloudiness, large-scale moisture throughout a deep layer, and large-scale, upper-level (~200 hPa) divergence, especially within the trough and southerly phases, suggesting that these parameters are most important for cyclogenesis and for distinguishing DWs from NDWs.

1. Introduction

Tropical easterly waves, including African easterly waves, are important for tropical cyclogenesis in the Atlantic (e.g., Landsea 1993) and in the east Pacific (e.g., Avila 1991; Avila and Pasch 1992; Molinari and Vollaro 2000). In particular, Hopsch et al. (2010) examined the possible relation between easterly wave structure near the West African coast and tropical cyclogenesis over the Atlantic. They used a wave-following Lagrangian framework for their analysis by creating composites of developing waves (DWs) and nondeveloping waves (NDWs) as a function of day relative to the day each wave trough moved over the West African coast. Results from Hopsch et al. (2010) indicated that DWs are associated with stronger mid- and low-level circulations and higher values of relative humidity compared to NDWs. They also found that DWs tend to begin with a cold-core structure far from the coast over Africa and develop a warm-core structure at the coast and over the ocean, consistent with cyclogenesis, while NDWs maintain a cold-core structure.

In addition to easterly wave structure, the environment through which a wave propagates may be another important influence on whether a wave spawns a tropical cyclone. Waves are more likely to develop tropical
cyclones when they move through environments characterized by higher SSTs, small wind shear, high moisture values, above-normal low-level relative vorticity, convergence at low levels, and divergence at upper levels (all conditions well known to be favorable for tropical cyclogenesis; e.g., Gray 1968; McBride and Zehr 1981; Landsea et al. 1998; Agudelo et al. 2011). In addition, Agudelo et al. (2011) found that waves were more likely to develop cyclones when they were convectively coupled and/or moved through environments with preexisting convection/diabatic heating, consistent with several other studies (e.g., Chronis et al. 2007; Hopsch et al. 2010; Leppert and Petersen 2010), which showed that DWs are associated with more widespread and/or vigorous convection relative to NDWs.

More widespread and/or intense convection may aid tropical cyclogenesis in several ways. For example, convection could act to moisten mid- and upper levels as a result of transport in convective updrafts. By inhibiting strong downdrafts and reducing the negative effects of entrainment, the increase of mid- and upper-level moisture could help to intensify subsequent updrafts and convection (e.g., Nolan 2007). This convection could also aid the intensification of a midlevel vortex (present, e.g., in the trough phase of an easterly wave) via the stretching term in the vorticity equation, tilting of horizontal vorticity into the vertical, and/or vertical advection (e.g., Trier et al. 1997; Arnault and Roux 2010). The increase in intensity and inertial stability of the midlevel vortex and resulting concentration of moisture and latent heating enhances conditions favorable for further convection, aiding the development of a smaller-scale, low-level vortex and tropical cyclogenesis [e.g., Bister and Emanuel 1997; Nolan 2007; the “marsupial paradigm” for tropical cyclogenesis developed by Dunkerton et al. (2009) also stresses the importance of moisture accumulation for cyclogenesis, in particular within a Lagrangian recirculation region within an easterly wave]. Note that the results of Nolan (2007) are based on idealized numerical simulations, but Raymond et al. (2011) found a similar process for cyclogenesis (i.e., the initial development of a midlevel vortex is favorable for enhancing convection, which subsequently aids the development of a low-level vortex) based on observations. Hence, convection may aid cyclogenesis by moistening the larger-scale environment and/or leading to the development and/or intensification of mid- to low-level vorticity.

Leppert et al. (2013) utilized data from the Tropical Rainfall Measuring Mission (TRMM) Lightning Imaging Sensor (LIS), Precipitation Radar (PR), and Microwave Imager (TMI) as well as IR brightness temperatures from the National Aeronautics and Space Administration (NASA) global-merged dataset to determine which specific characteristics/observations of convection were most important for tropical cyclogenesis and distinguishing DWs from NDWs over fixed longitude bands (i.e., Eulerian framework). They found that as waves neared their genesis location over the Atlantic, the coverage by IR brightness temperatures ≤240 and ≤210 K (i.e., cold cloudiness coverage) provided the greatest distinction between DWs and NDWs. In contrast, lightning flash rates, low-level (i.e., below 3.5 km) PR convective reflectivity (i.e., indicators of convective intensity), and convective coverage in addition to the coverage by IR brightness temperatures below certain thresholds provided a statistically significant distinction only between east Pacific DWs (i.e., waves that developed a tropical cyclone over the east Pacific) and NDWs as east Pacific DWs approach their development region. Thus, results from Leppert et al. (2013) suggest that the coverage by cold cloud tops and convection in precursor easterly waves is more important for tropical cyclogenesis over the Atlantic than the intensity of convection, while the coverage and intensity may both be important for genesis over the east Pacific.

The results from Leppert et al. (2013) using an Eulerian framework may have forecasting applications by helping to distinguish DWs from NDWs, but this framework cannot provide much insight into the evolution of waves in the days leading up to tropical cyclogenesis. Therefore, this study seeks to build upon Leppert et al. (2013) by using a Lagrangian methodology to obtain a greater understanding of the genesis process.

Specifically, this study has two goals. The first is to determine the evolution of several convective characteristics (e.g., coverage and intensity) and the corresponding evolution of the thermodynamics and dynamics of larger-scale easterly waves in the days leading up to and including tropical cyclogenesis. Second, we seek to determine whether this evolution changes for easterly waves that develop cyclones over different regions. The first hypothesis is that convective coverage and/or intensity increases as genesis is approached for DWs, and these increases correspond with the development of the vortex and warm core. We hypothesize in this work that such a trend in coverage and/or intensity can help distinguish DWs from NDWs. In the next section, a description of the methodology is provided. Section 3 provides the main results from the study, supporting the first hypothesis but not the second. The summary and conclusions are given in section 4.
2. Methodology

a. Easterly wave identification

Using the same methodology as Leppert and Petersen (2010) and Leppert et al. (2013), easterly waves were analyzed and separated into ridge, northerly, trough, and southerly phases using National Centers for Environmental Prediction–National Center for Atmospheric Research (NCEP–NCAR) reanalysis (Kalnay et al. 1996) 700-hPa meridional wind data [note that the reanalysis has a spatial (temporal) resolution of 2.5° (6 h; averaged to 1 day for this study)] for June–November of 2001–10 over a region stretching 5°–20°N, 130°W–20°E (Fig. 1) from the east Pacific to West Africa. Specifically, wave phases were identified by first calculating a daily average meridional wind value between 5° and 20°N (i.e., wind is not allowed to vary with latitude). Then a 3–7-day bandpass filter was applied to meridional wind anomalies calculated relative to the mean at each longitude. The filtered anomalies were subsequently normalized by the standard deviation valid at each longitude, and the ±0.75 standard deviation threshold was used to identify the individual wave phases, as described in Leppert and Petersen (2010) and Leppert et al. (2013). The identification of wave phases using only meridional wind left many data points unable to be classified as any wave phase. Therefore, as in Leppert and Petersen (2010) and Leppert et al. (2013), 700-hPa vorticity calculated using reanalysis zonal and meridional wind components was processed exactly as the meridional wind data. Then the resulting normalized, filtered vorticity anomalies were used to identify those wave phases unable to be classified with the meridional wind data alone using a similar ±0.75 standard deviation threshold, although some data points could still not be classified and were not used in this study.

After the various wave phases were identified, information from the National Hurricane Center (NHC) storm reports (National Hurricane Center 2011) was utilized to identify easterly wave troughs that were associated with the development of a tropical cyclone of at least tropical storm strength (i.e., DWs). National Hurricane Center (2011) was also used to determine the day on which a tropical depression was first identified (i.e., day zero; D0) for each DW. Then each wave trough was traced up to 5 days prior to D0 (D − 5) until 1 day after D0 (D + 1). Any of the other three wave phases found within three data points (7.5°) east or west of the DW troughs were considered to be part of the DW.

In addition, we sought to compare the evolution of DWs with that of waves that never developed a tropical cyclone (i.e., NDWs). Day zero for NDWs was defined as the day on which these waves achieved their maximum 850-hPa vorticity within the trough phase. Tropical cyclogenesis requires a finite-amplitude, low-level disturbance (e.g., Kurihara and Tuleya 1981; Rotunno and Emanuel 1987; Emanuel 1989), and a strong, coherent NDW is more likely to provide this low-level disturbance than a NDW with a weak and/or short-lived low-level circulation. Therefore, the strongest, most coherent NDWs are presumably most relevant for comparison to DWs. Hence, only those NDWs that could be tracked for at least seven days (NDWs were tracked manually using Hovmöller diagrams) and achieved a maximum 850-hPa vorticity of at least $8.0 \times 10^{-6}$ s$^{-1}$ (equal to the mean 850-hPa vorticity of all NDW troughs plus one standard deviation) were included in the analysis. Note that the mean 850-hPa vorticity of DWs on D0 ($5.1 \times 10^{-6}$ s$^{-1}$) is less than the NDW threshold. It is possible that at least some of the NDWs included here were associated with especially strong equatorward advection of dry air which may have inhibited their development of tropical cyclones. In other words, some of the NDWs included here may have been too strong to develop cyclones (Hopsch et al. 2010). Once D0 was identified for each NDW trough, D − 5 through D + 1 as well as the other NDW phases were identified similar to DWs.

As an example of the wave tracking procedure relative to D0, Fig. 2 shows a time–longitude plot of the normalized, filtered, 700-hPa meridional wind anomalies for 2010 and the locations of DW and NDW troughs.

![Fig. 1. Map showing the location of the analysis domain utilized for this study as indicated by the black outline.](http://journals.ametsoc.org/mwr/article-pdf/141/8/2649/4297695/mwr-d-12-00217_1.pdf)
valid for $D - 5$ through $D + 1$ in 2010 identified using both meridional wind and vorticity information. The “×” symbols in Fig. 2 mark the location of initial tropical depression identification (i.e., D0) by the NHC for each DW. A comparison between the shaded regions (representing meridional wind) and the small squares/triangles (representing wave troughs identified using information from both meridional wind and vorticity) indicates that waves are generally easier to track using meridional wind and vorticity information as opposed to using meridional wind only. However, some waves are still difficult to track even with the combined information (e.g., wave which spawned east Pacific Hurricane Darby on 23 June at 92.9°W).

b. Composite methodology

Using a similar methodology to that of Hopsch et al. (2010), composites of vertical motion and specific humidity from the NCEP–NCAR reanalysis were created as a function of wave phase and day relative to D0 for DWs and NDWs. In addition, divergence and vorticity were calculated by applying fourth-order finite-difference methods to reanalysis meridional and zonal wind components, and equivalent potential temperature $\theta_e$ was calculated using Eq. (38) from Bolton (1980). Then divergence, vorticity, and $\theta_e$ were composited similar to omega and specific humidity to examine the evolution of the structure/circulation of synoptic-scale easterly waves. Note that the resolution/accuracy of the reanalysis is limited. Hence, we only use it to identify general characteristics of easterly waves.

To assess the evolution of the coverage by cold cloudiness associated with easterly waves, the fractional coverage by IR brightness temperatures $\leq 240$ and $\leq 210$ K were calculated using data from the NASA global-merged IR brightness temperature dataset (Liu et al. 2009). In addition, the convective/stratiform classification (Awaka et al. 1998, 2009) contained in the TRMM PR 2A25 V6.0 product was used to calculate the percentage of PR pixels in each 2.5° box classified as convective (i.e., percentage convective coverage). Both the convective and IR threshold coverage were subsequently composited as a function of wave phase and day relative to D0.

Three parameters were used to analyze the evolution of convective intensity. Lightning data from the LIS (Christian et al. 1992; Boccippio et al. 2000, 2002) were used to calculate flash rates valid for each day and 2.5° grid box. Deep convective updrafts and robust mixed-phase microphysical processes (i.e., intense convection) have been shown to be a prerequisite for the development of strong in-cloud electric fields and associated lightning (e.g., Takahashi 1978; Rutledge et al. 1992; Williams et al. 1992; Zipser 1994; Saunders and Peck 1998; Deierling and Petersen 2008). In addition, previous studies (e.g., Chronis et al. 2007; Price et al. 2007) have shown that enhanced lightning may be related to tropical cyclogenesis over the east Atlantic. Brightness temperatures measured at 37.0 and 85.5 GHz from the TMI (Kummerow et al. 1998) were also used to infer
information about convective intensity. The measured radiances at these frequencies are especially sensitive to scattering by ice (e.g., Spencer et al. 1989; Smith et al. 1992; Cecil and Zipser 1999; Toracinta et al. 2002). Therefore, a significant reduction in measured brightness temperatures at 37.0 and/or 85.5 GHz likely indicates the presence of large ice particles that require strong convective updrafts to form and be held aloft in the upper portions of clouds. In particular, the brightness temperatures measured at both polarizations of the 85.5-GHz frequency were combined to form 85.5-GHz polarization corrected temperatures (PCT85), and the two 37.0-GHz channels were combined into PCT37. Only those TMI pixels with PCT85 ≤ 200 K and PCT37 ≤ 260 K were used in the calculation of average PCT85 and PCT37 over each 2.5° box. This results in mean PCT values that are indicative of how strong the convection is, whereas a mean PCT without any such thresholds would be more responsive to the coverage by deep precipitation. Also used to assess convective intensity was attenuation-corrected radar reflectivity data (Iguchi et al. 2000; Meneghini et al. 2000; Iguchi et al. 2009) from the PR (Kummerow et al. 1998; Kozu et al. 2001). Mean profiles of convective reflectivity (using only data classified as convective with a rain bottom below 2 km and not classified as warm rain) were calculated for each 2.5° box stretching from 1–18 km above mean sea level with 1-km resolution. These mean profiles in addition to lightning flash rates and mean PCTs were composited as a function of day relative to D0 and wave phase, similar to the reanalysis variables and coverage parameters. See Leppert et al. (2013) for more detail on the three convective intensity parameters.

c. Statistical analysis

The analysis of variance statistical technique was utilized to determine whether the systematic variation between DWs and NDWs was statistically significantly greater than the random variation of the sample of data (i.e., whether differences between the DW and NDW categories were statistically significant). See Panofsky and Brier (1958) for more information on the analysis of variance technique.

3. Results

a. Composite analysis

The DW [225 distinct easterly waves of which 122 (103) developed a cyclone over the east Pacific (Atlantic)] and NDW [180 distinct easterly waves of which 20 (160) achieved their maximum 850-hPa vorticity over the east Pacific (Atlantic)] composites incorporate waves that developed a tropical cyclone and/or achieved maximum 850-hPa vorticity anywhere in the full 130°W–20°E analysis domain, including near the West African coast and the west coast of Central America. Hence, the evolution of some easterly waves used in the composites from D − 5 to D0 includes a transition from over land to ocean. Because convective intensity and lightning generally decrease over the ocean compared to land (e.g., Zipser 1994; Toracinta et al. 2002; Christian et al. 2003), it is likely that a decrease in the intensity of convection may be observed for some waves as they evolve from D − 5 to D0 unrelated to the evolution toward cyclogenesis and/or maximum low-level vorticity. Therefore, a land mask was applied to the DW and NDW composites. Table 1 provides the number of individual data points used for the composites of these two wave categories after application of the land mask. Although circulations associated with the ridge phase may influence tropical cyclogenesis in the trough phase (e.g., Arnault and Roux 2010, 2011), cyclogenesis is not expected to occur in the ridge phase, which is associated with a midlevel anticyclonic circulation. In addition, the ridge sample sizes (Table 1) are much smaller than those of the other phases. Hence, ridge phase composites are not shown.

1) CONVECTIVE COVERAGE VARIABLES

The composite coverage by certain IR brightness temperature thresholds is shown in Table 2 for DWs and NDWs (as a function of wave phase and day relative to D0). Statistically significant 2-day changes in coverage are also identified in Table 2. Two-day changes allow for gradual changes to become more clearly established and
reduce the influence of short-term variability. The values in Table 2 for DWs indicate that the trends in coverage by cold cloudiness vary as a function of phase in the days leading up to tropical cyclogenesis. For example, the coverage by temperatures \( \leq 240 \) K significantly increases from \( D - 5 \) until \( D - 2 \) in the northerly phase and then decreases thereafter. In contrast, the coverage of the 240-K threshold significantly increases from \( D - 5 \) through D0 in the trough and southerly phases (note that coverage decreases on \( D + 1 \) in these wave phases). The trends shown by the coverage of the 210-K threshold indicate an initial increase by coverage in the northerly phase up to \( D - 2 \) followed by a decrease. The coverage by deep, cold cloud tops (i.e., temperatures \( \leq 210 \) K) more than doubles between \( D - 5 \) and D0 in both the trough and southerly phases. In summary, cold cloudiness increases in the northerly, trough, and southerly phases a few days before genesis and continues to increase through genesis in the trough and southerly phases. Thus, near D0, cold cloudiness becomes predominant in the trough and southerly phases consistent with Kiladis et al. (2006) found for all easterly waves (not just DWs) as they move from over Africa to the east Atlantic.

The values for NDWs in Table 2 suggest an evolution in the coverage by cold cloudiness that is similar to that of DWs. In particular, all phases show a general increasing trend across all days except \( D + 1 \). While the evolution of coverage by cold cloud tops is similar between DWs and NDWs, the magnitude of values are significantly greater (i.e., statistical significance at or above the 99% level) for DWs on nearly all days in all phases. Hence, it appears that a greater coverage by cold cloudiness is more relevant to tropical cyclogenesis than the evolution of that coverage in the days leading up to genesis.

The pattern observed in the evolution of convective coverage for DWs (Fig. 3a) is somewhat similar to that observed for the IR threshold coverage associated with those waves. Only the trough and southerly phases have much increase in convective coverage from \( D - 5 \) to D0, but the 2-day changes are not significant, except for the increase from \( D - 3 \) to \( D - 1 \) in the southerly phase. The coverage by IR brightness temperatures less than certain thresholds shown in Table 2 includes not only cold cloudiness directly over active convection but also the divergent anvil cloud. In contrast, the percentage convective coverage shown in Fig. 3 includes only the coverage by active convection as indicated by the PR convective-stratiform algorithm. Thus, the faster increase in the coverage by cold cloudiness (Table 2) than the coverage by convection (Fig. 3) as D0 is approached in the DW trough and southerly phases suggests that the anvil cloud is expanding faster than convection. This relatively rapid increase in coverage by cold cloudiness

### Table 2

<table>
<thead>
<tr>
<th>Day</th>
<th>Developing waves</th>
<th>Nondeveloping waves</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Northerly Trough</td>
<td>Southerly Northerly Trough Southerly</td>
</tr>
<tr>
<td>240-K threshold</td>
<td></td>
<td></td>
</tr>
<tr>
<td>( D - 5 )</td>
<td>0.109</td>
<td>0.112</td>
</tr>
<tr>
<td>( D - 4 )</td>
<td>0.107</td>
<td>0.122</td>
</tr>
<tr>
<td>( D - 3 )</td>
<td>0.124</td>
<td>0.121</td>
</tr>
<tr>
<td>( D - 2 )</td>
<td>0.137</td>
<td>0.146</td>
</tr>
<tr>
<td>( D - 1 )</td>
<td>0.132</td>
<td>0.152</td>
</tr>
<tr>
<td>D0</td>
<td>0.114</td>
<td>0.171</td>
</tr>
<tr>
<td>( D + 1 )</td>
<td>0.100</td>
<td>0.163</td>
</tr>
<tr>
<td>210-K threshold</td>
<td></td>
<td></td>
</tr>
<tr>
<td>( D - 5 )</td>
<td>0.011</td>
<td>0.014</td>
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<tr>
<td>( D - 4 )</td>
<td>0.013</td>
<td>0.017</td>
</tr>
<tr>
<td>( D - 3 )</td>
<td>0.017</td>
<td>0.017</td>
</tr>
<tr>
<td>( D - 2 )</td>
<td>0.019</td>
<td>0.021</td>
</tr>
<tr>
<td>( D - 1 )</td>
<td>0.019</td>
<td>0.022</td>
</tr>
<tr>
<td>D0</td>
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<td>0.029</td>
</tr>
<tr>
<td>( D + 1 )</td>
<td>0.014</td>
<td>0.030</td>
</tr>
</tbody>
</table>

**Fig. 3.** Percentage convective coverage as a function of day relative to D0 for various wave phases of (a) developing waves and (b) nondeveloping waves after application of a land mask. The thin, horizontal line indicates the average convective coverage over the full analysis domain, and the standard deviation is 1.78%.
may suggest an increase in larger-scale, upper-level divergence in the trough and southerly phases, as will be shown later.

The evolution of convective coverage for NDWs (Fig. 3b) generally does not show any consistent trends or significant 2-day changes. In addition, a comparison between the two plots shown in Fig. 3 indicates that convective coverage is generally greater for DWs, but the analysis of variance statistical technique shows that the differences are not significant, except for the southerly phase values near D0.

2) CONVECTIVE INTENSITY VARIABLES

In contrast to what is observed for the coverage by convection/cold cloudiness (especially in the trough and southerly phases), the intensity of DW convection as indicated by trends in lightning (Fig. 4a) generally appears to decrease as D0 is approached. Specifically, both the trough and northerly phases show a decrease in flash rate every day after D = 4, except the trough phase on D0. The DW southerly phase shows a decrease in flash rate from D = 5 to D = 2, an increase from D = 2 to D0, and a relatively large decrease on D + 1. However, the DW southerly phase flash rate on D + 1 is less than that on D = 5, suggesting an overall decrease in the intensity of convection with time for that wave phase. Note that trends observed in any of the DW phases are small compared to the standard deviation of the flash rate of 1133.3 day\(^{-1}\) (0.5\(^{\circ}\))\(^{-2}\) but generally larger than the 25th and 75th percentile values of 0.0 and 2.5 day\(^{-1}\) (0.5\(^{\circ}\))\(^{-2}\), respectively. The flash rates shown in Fig. 4 are valid over only water, but lightning flash rates are generally higher over water near the coast than over ocean farther from land (cf. Fig. 4 from Christian et al. 2003). To account for this effect, we applied a land mask prior to creating the lightning composites in which all data points classified as water within one data point (2.5\(^{\circ}\)) of land were reclassified as land and excluded from the composites, and another mask was used where points classified as water within two data points (5.0\(^{\circ}\)) of land were excluded. Results from both adjusted land masks (not shown) reveal similar decreasing trends in flash rate as D0 is approached. Therefore, the decrease in flash rate in all phases is not due to the propagation of waves from over land to water or from near the coast to open water.

The LIS flash rates of NDWs (Fig. 4b) show little indication of any consistent trends from D = 5 through D0. However, D0 and D + 1 values in the NDW northerly and trough phases (southerly phase) are greater than the D − 5 (D − 3) value of those same wave phases, suggesting an increase in the intensity of convection for NDWs with time. It is also possible that more flashes are recorded as a result of the greater coverage by potentially electrically active cloud, while the flash rate (i.e., intensity) of individual convective elements does not change. This contrasts with DWs, which show a decrease in lightning with time despite an increase in coverage of cold cloudiness. An examination of the distribution of TRMM observations (including LIS) as a function of time of day (not shown) indicates that waves were sampled nearly uniformly over the diurnal cycle, including the NDW southerly phase on D = 4. Hence, it is not clear why the flash rate on D = 4 in the NDW southerly phase is so large [i.e., 907.4 day\(^{-1}\) (0.5\(^{\circ}\))\(^{-2}\)]. Perhaps the LIS happened to sample this wave phase on D = 4 preferentially when convection was especially electrically active.

A comparison between Figs. 4a and 4b shows that over most days, the difference in lightning flash rates between DWs and NDWs is small with some values greater for DWs and others greater for NDWs. However, closer to D0, flash rates generally become greater for NDWs. In fact, the flash rate in the NDW trough on D − 1 is significantly greater than the corresponding DW value. While the NDW southerly phase flash rate on D + 1 is not significantly greater than the corresponding DW value at the 99% level, it is significantly greater at the 95% level.

Figure 5 shows the evolution of composite PCTs for DW and NDW phases that generally suggest trends in
convective intensity consistent with those inferred from the evolution of lightning flash rates. Both the PCT$_{37}$ and PCT$_{85}$ for DWs display much variability from day to day, but the overall trend for all phases at both frequencies from $D - 5$ to $D + 1$ is positive (i.e., a weaker ice scattering signature with time). These positive PCT trends suggest a negative trend in the amount and/or size of graupel produced and supported by convective updrafts and an associated decrease in the intensity of convection as the waves evolve toward cyclogenesis. NDWs also show much variability in PCTs at both frequencies with little evidence of consistent trends, especially for PCT$_{37}$ values. However, the PCT$_{85}$ values do show a decrease in magnitude from $D - 5$ until D0 and $D + 1$ in all phases. Overall, the NDW PCTs suggest an intensity of convection that changes little with time. A comparison between DW and NDW PCTs shows that differences between the two wave types are generally small (i.e., not statistically significant at the 99% level). However, on D0, PCT$_{85}$ and PCT$_{37}$ values are significantly

![composite graphs showing PCT values for DWs and NDWs](image-url)
greater (i.e., weaker convection) in all DW phases. DW PCT\textsubscript{85} values are also significantly greater than NDW values in the trough and southerly phases on D – 1. Thus, both the Eulerian analysis of Leppert et al. (2013) and the Lagrangian analysis presented here reveal few statistically significant differences between DWs and NDWs using mean cold PCT\textsubscript{85} and PCT\textsubscript{37} (i.e., PCT\textsubscript{85} ≤ 200 K and PCT\textsubscript{37} ≤ 260 K). However, the analysis presented here does suggest a significantly weaker ice scattering signature for DWs on the day of genesis.

Another indicator of convective intensity is mean profiles of PR convective reflectivity shown as a function of wave phase and day relative to D0 for DWs and NDWs in Fig. 6. The DW profiles change little with time, while NDWs show a small but noticeable increase in reflectivity from D = 5 to D0 between ~2.5 and ~6.5 km in all phases. These differences for NDWs between D = 5 and D0 are actually significant at the 99% level near 4.5 km in all phases. This increase in NDW reflectivity with time suggests an increase in the intensity of updrafts and convection with time in order to support and produce the larger reflectivity values. A comparison between DW and NDW reflectivity values indicates that values in all phases are significantly greater for DWs in the ~2.5–4.5-km layer on D = 5 and D = 3. After D = 3, differences between the two wave categories are generally not significant, though some phases at scattered heights show significantly smaller values for DWs (e.g., 6.5-km trough value on D = 1). Thus, no discernible trends in convective intensity can be inferred for DWs from mean PR convective reflectivity profiles, in contrast to lightning flash rates (Fig. 4) and PCTs (Fig. 5), which suggest a decrease in convective intensity with time. The PR convective reflectivity profiles for NDWs suggest a slight increase in convective intensity with time, consistent with lightning and PCT information.

Overall, differences in convective reflectivity between DWs and NDWs are generally small. Vertical profiles of the 90th percentile of reflectivity values (not shown) were also examined to better understand the evolution of the largest reflectivity values. In general, the evolution of these profiles is similar to that of the mean profiles.

Various composites were also created for waves that spawned tropical cyclones and/or achieved maximum low-level vorticity over the separate longitude bands identified in Fig. 1 of Leppert et al. (2013). In general, patterns of convective coverage/intensity observed for these waves over the smaller longitude bands were similar to those shown for waves over the full analysis domain. However, the evolution of mean convective reflectivity profiles valid for waves that developed a tropical depression over the west Atlantic (i.e., 40\degree–70\degreeW; not shown) and over the western Caribbean (between the east coast of Central America and 70\degreeW; Fig. 7) is different from that observed for DWs over the full analysis domain (Figs. 6a,c,e). In particular, the Caribbean DW trough and southerly phases are associated with clear increases in reflectivity from D = 5 to D0 below 6.5 km (increase in the 2.5–5.5-km layer of both phases is significant at the 99% level), while virtually no change is seen for DWs valid over the full domain in any wave phase. Thus, the intensity of convection appears to increase as D0 is approached for Caribbean DWs, at least in some wave phases, while the intensity of DWs over the full domain appears to remain approximately constant with time in all phases as indicated by reflectivity information. This suggests that convective intensity may be relatively more important for tropical cyclogenesis over certain regions, similar to the results of Leppert et al. (2013).

3) LARGE-SCALE VARIABLES

Despite a decrease in intensity with time (as indicated by lightning flash rates and PCTs), the increasing coverage of convection and cold cloudiness in the DW trough and southerly phases valid over the full domain may influence the larger-scale easterly wave as a result of a larger net latent heating. This heating may cause the heights of pressure surfaces below (above) the heating to decrease (increase). The resulting adjustments of the momentum field may then lead to convergence (divergence) below (above) the heating. Figure 8 shows vertical profiles of divergence for various DW and NDW phases as a function of day relative to D0. None of the DW wave phases show much change in low-level (below 850 hPa) convergence, and the northerly phase shows virtually no change in upper-level (~200 hPa) divergence from D = 5 to D0. However, the DW trough and southerly phases (phases that exhibit an increase in coverage of cold cloudiness and convection) are associated with increases in upper-level divergence as genesis is approached. The increase is relatively small in the trough phase (increase of \(-1.5 \times 10^{-6} \text{ s}^{-1}\) from D = 5 to D0), but divergence in the southerly phase doubles from D = 5 to D0.

The evolution shown by NDW profiles of divergence (Figs. 8b,d,f) is somewhat different than that observed for DWs. In particular, low-level convergence increases in all NDW phases from D = 5 to D0. At earlier days, NDW convergence is less than the corresponding, nearly constant DW values, but by D = 1, the composite low-level convergence is similar between DWs and NDWs. In addition, upper-level divergence in all NDW phases increases with time (perhaps related to the increase in
Fig. 6. Composite analysis of vertical profiles of precipitation radar convective reflectivity (using only values classified as convective) as a function of day relative to D0 for the (a) developing wave (DW) northerly, (b) non-developing wave (NDW) northerly, (c) DW trough, (d) NDW trough, (e) DW southerly, and (f) NDW southerly phases (a land mask has been applied). The $D-4$ and $D-2$ profiles are not shown for clarity, and the horizontal dashed lines show the value of 18 dBZ plus one standard deviation at each height.
cold cloudiness coverage observed in all those wave phases; Table 2), whereas this behavior is only seen in the DW trough and southerly phases. A comparison between DW and NDW values in Fig. 8 also shows that upper-level divergence in the DW trough and southerly phases is generally larger than the corresponding NDW values (many values are significantly larger, valid at the 99% level).

A difference between the evolution of DWs valid over the full analysis domain and that of DWs over smaller longitude bands can be seen for the evolution of divergence profiles. The vertical profiles of divergence for Caribbean DWs (Fig. 9) show an increase in convergence below 850 hPa for all phases between D = 5 and D0, in contrast to DWs over the full domain (Figs. 8a,c,e) which displayed virtually no change in low-level convergence in any wave phase. It is possible that the increase in convective intensity inferred for Caribbean DWs as they evolve toward cyclogenesis is associated with an increase in the depth and magnitude of latent heating. This increase in heating may exert a larger influence on low-level mass and momentum fields, resulting in stronger Caribbean DW low-level convergence on scales that can be resolved by the NCEP–NCAR reanalysis dataset. This large-scale convergence may aid the merger of any convective-scale cyclonic vorticity anomalies created by the updrafts of intense convection, aiding the increase of larger-scale, low-level vorticity (e.g., Hendricks et al. 2004; Montgomery et al. 2006) and tropical cyclogenesis.

Upper-level divergence should be associated with greater upward motion below the divergence due to mass conservation. The vertical profiles of omega shown for DWs valid over the full domain in Figs. 10a,c,e, indeed, show an increase in upward vertical motion with time below ~200 hPa (with a maximum increase ~300–400 hPa) in the trough and southerly phases, consistent with the increase in upper-level divergence in those wave phases. The vertical profiles of omega for NDWs in Figs. 10b,d,f also show an increase in upward vertical motion with time throughout a large depth of the troposphere in all phases, consistent with the increase in upper-level (low-level) divergence (convergence) observed for those waves. The upward motion is generally significantly greater in the DW trough and southerly phases through a large depth of the troposphere compared to the corresponding NDW values, except in the trough on D0.

Another potential feedback of convection to the larger scale that could aid cyclogenesis may be the moistening of mid- and upper levels (e.g., Rotunno and Emanuel 1987; Nolan 2007; Dunkerton et al. 2009), leading to an increase in $\theta_e$. The evolution of $\theta_e$ anomaly profiles

![Fig. 7. Composite analysis of vertical profiles of precipitation radar convective reflectivity (using only values classified as convective) as a function of day relative to D0 for the (a) northerly, (b) trough, and (c) southerly phases of Caribbean developing waves (CDWs; a land mask has been applied). The D = 4 and D = 2 profiles are not shown for clarity, and the horizontal dashed lines show the value of 18 dBZ plus one standard deviation at each height.](http://journals.ametsoc.org/mwr/article-pdf/141/8/2649/4297695/mwr-d-12-00217_1.pdf)
FIG. 8. Composite analysis of vertical profiles of divergence as a function of day relative to D0 for the (a) DW northerly, (b) NDW northerly, (c) DW trough, (d) NDW trough, (e) DW southerly, and (f) NDW southerly phases (a land mask has been applied). The $D - 4$ and $D - 2$ profiles are not shown for clarity, and the horizontal dashed lines indicate ± one standard deviation at each pressure level.
shown in Fig. 11 for DWs and NDWs indicates an increase in all DW phases throughout the depth of the troposphere from \( D = 5 \) to \( D + 1 \), but the increases are generally more pronounced in those wave phases that show a consistent positive trend in the coverage by cold cloudiness/convection (i.e., trough and southerly phases). In contrast, the corresponding NDW \( \theta_e \) anomaly profiles generally indicate either little change with time or no consistent trends from \( D = 5 \) to \( D + 1 \). In addition, a comparison between corresponding DW and NDW profiles reveals that DWs are associated with significantly more moisture than NDWs (in fact, many \( \theta_e \) anomaly values are significantly greater for DWs valid at the 99% level, especially after \( D = 5 \)), consistent with what Hopsch et al. (2010) found for easterly waves near the West African coast.

The evolution of profiles of relative vorticity valid for DWs and NDWs is shown in Fig. 12 as a function of wave phase. The DW northerly and southerly phases show little to no change with time in vorticity below 600 hPa. However, the trough (phase where cyclogenesis is assumed to occur) does show an increase in cyclonic vorticity at midlevels (~500–600 hPa) after \( D = 3 \) and at low levels (below 850 hPa) after \( D = 1 \) (although, these increases are admittedly small). Hence, it appears that midlevel vorticity in the trough increases before that at low levels, consistent with the numerical simulations of Nolan (2007). It is not possible to determine from Fig. 12c whether the initial increase at midlevels subsequently develops downward (i.e., top-down genesis mechanism; e.g., Bister and Emanuel 1997). Because the top-down and bottom-up [where vorticity initially increases at low levels and develops upward; e.g., Hendricks et al. (2004); Montgomery et al. (2006); Houze et al. (2009)] genesis mechanisms generally occur on meso- and smaller scales, the resolution of the reanalysis used is unable to robustly resolve these mechanisms. At upper levels (~200 hPa), the DW trough phase exhibits an increase in anticyclonic vorticity with time. This upper-level increase combined with the increase in mid- and low-level cyclonic vorticity suggests the presence of a warm core in association with thermal wind balance, consistent with tropical cyclogenesis. Hopsch et al. (2010) also found that a warm core structure develops in association with DWs over Africa as they move toward the east Atlantic, closer to tropical cyclogenesis.

The evolution of NDW vorticity profiles shown in Fig. 12 indicates that all NDW phases, except the trough, are generally associated with little change with time and/or no consistent trends. In contrast, the NDW trough phase shows an increase in vorticity below 400 hPa on \( D = 1 \) and \( D_0 \) followed by a decrease on \( D + 1 \). Mid- to low-level vorticity continued to increase.
slightly on $D + 1$ for DWs (Fig. 12c). The peak 850-hPa vorticity in the NDW trough on D0 is a result of how D0 is defined for NDWs. The NDW trough also exhibits an increase in upper-level anticyclonic vorticity until $D - 1$ that remains approximately constant thereafter. However, a comparison between Figs. 12c and 12d shows that upper-level anticyclonic vorticity is larger on all days for the DW trough.
Fig. 11. Composite analysis of vertical profiles of the $\theta_e$ anomaly (anomaly relative to the mean at each pressure level and longitude) as a function of day relative to D0 for the (a) DW northerly, (b) NDW northerly, (c) DW trough, (d) NDW trough, (e) DW southerly, and (f) NDW southerly phases (a land mask has been applied). The $D - 4$ and $D - 2$ profiles are not shown for clarity, and the horizontal dashed lines indicate the value of the standard deviation at each pressure level.
In summary, the coverage by cold cloudiness/convection increases, while convective intensity appears to decrease as genesis (i.e., D0) is approached for DWs. The coverage by cold cloudiness also increases with time for NDWs, but values are generally larger for DWs. The convective intensity of NDWs appears to increase slightly or remain approximately constant with time. However, an increase in the intensity of convection
and 850-hPa vorticity (as observed for NDWs on D0) is apparently not sufficient for cyclogenesis. Perhaps the development of favorable upper-level conditions (i.e., increased upper-level divergence, anticyclonic vorticity, and upward motion) and greater moisture throughout a large depth of the troposphere in association with greater coverage by convection and cold cloudiness as seen for DWs (especially in the trough) are more important for cyclogenesis.

b. Conceptual model

The above analysis can be synthesized into a conceptual model describing the evolution of convection, cold cloudiness, and large-scale easterly waves in the days leading up to tropical cyclogenesis. From $D - 5$ up to cyclogenesis, the coverage by convection and cold cloudiness increases (as shown in the transition from part a to part b of the schematic depicted in Fig. 13) in the trough and southerly phases, helping to moisten the larger-scale environment throughout a large depth of the troposphere (Figs. 11c,e). This increase in moisture is consistent with what Nolan (2007) found prior to tropical cyclogenesis and may help to inhibit strong, evaporatively cooled downdrafts (Rotunno and Emanuel 1987), allowing for more persistent convection. Arnault and Roux (2011) found that a necessary condition for cyclogenesis near the West African coast was deep, sustained convection in the easterly wave trough. Persistent cold cloudiness and convection may allow sufficient time for the development of a mesoscale convective vortex at midlevels (Fig. 13b) and an associated increase in midlevel vorticity (as observed in the trough phase in Fig. 12c). The increase in vorticity and associated increase in inertial stability may help to concentrate latent heating (i.e., limit its transport away from the developing circulation and tropical cyclone), thus allowing the heating to exert a greater impact on the larger scale as discussed below.

The inhibition of strong downdrafts may also lead to the reduction of low-level cooling and the development of a more convective-type heating profile (e.g., Houze 1989). This type of profile combined with the increasing area and/or volume experiencing latent heating associated with the increase in coverage by cold cloudiness may aid the development of a transverse circulation (e.g., Shapiro and Willoughby 1982; Montgomery et al.)
Specifically, the heating may influence the mass field (Fig. 13b) so that the heights of isobaric levels above the heating increase while levels below the heating are forced downward slightly. The momentum field eventually adjusts to the changing mass distribution with increased upper-level divergence and smaller lower-level convergence. Note that the effect of a given amount of latent heating on the mass/momentum fields at low levels may be expected to be smaller than that at higher levels due to a higher density at lower levels and may help explain why an increase in low-level convergence is generally not observed in Figs. 8a,c,e. Alternatively, an increase in low-level convergence on a smaller scale than can be resolved by the reanalysis data may also help explain the low-level pattern observed in Figs. 8a,c,e. A pattern of enhanced upper-level divergence and smaller low-level convergence may then result in greater upward vertical motion (Fig. 13c). The combination of upward motion, divergence, and relatively small convergence (i.e., transverse circulation) may then act to remove more mass from above a developing tropical cyclone, leading to the development of a surface low pressure center and a surface cyclonic circulation (i.e., tropical cyclogenesis) via geostrophic adjustment (e.g., Arnault and Roux 2011). This conceptual model describes the initial development and/or intensification of a midlevel circulation followed by the development of a low-level circulation, but it is not clear given the limitations of the datasets used whether the mid- and low-level circulations develop independently (e.g., Nolan 2007) or the low-level circulation develops from the midlevel circulation (i.e., top-down mechanism).

4. Summary and conclusions

Tropical easterly waves are examined in this study via a wave-following Lagrangian framework to assess the evolution of convection and cold cloudiness as well as the evolution of larger-scale wave structure/circulation in the days leading up to and including tropical cyclogenesis (i.e., D0). In addition, D0 was defined for non-developing waves (NDWs) as the day on which these waves achieved a maximum in 850-hPa vorticity in order to facilitate a comparison between developing waves (DWs) and NDWs. This study also sought to determine whether the evolution of the two wave types varied over different regions.

Results suggest that the coverage by convection and cold cloudiness increases as genesis is approached for DWs valid over the full 130°W–20°E analysis domain, in particular within the trough and southerly phases. In contrast, the intensity of convection appears to decrease with time in all DW phases. The evolution of NDWs also generally shows an increase in coverage by cold cloudiness as D0 is approached, but convective intensity appears to remain approximately constant with time, perhaps increasing slightly. The biggest difference here between DWs and NDWs is that the coverage by cold cloudiness is generally significantly greater for DWs, consistent with the results of Leppert et al. (2013).

Consistent with the increase in cold cloudiness coverage, the DW trough and southerly phases exhibit an increase in upper-level (~200 hPa) divergence, upward vertical motion below this increase in divergence, and increased anticyclonic vorticity as D0 is approached. NDWs generally show similar patterns, but the increases in these parameters are smaller than DWs. These patterns shown by DWs and NDWs are generally consistent with the corresponding composites from Hopsch et al. (2010) as the waves move over the African coast. The DWs also show a small increase in mid- to low-level (below 500 hPa) cyclonic vorticity in the trough phase (phase presumably most important for cyclogenesis), while NDWs show little evolution in low-level vorticity except for a large increase in 850-hPa vorticity in the trough on D0 as a result of how D0 is defined for these waves. While the circulation associated with DWs and NDWs shows relatively small differences, the thermodynamic structure of the large-scale waves show more significant differences. In particular, \( \theta_e \) generally increases throughout a large depth of the troposphere for all DW phases as genesis is approached, and the \( \theta_e \) values on nearly all days and levels in all phases are significantly greater for DWs compared to NDWs. Therefore, it appears that a greater coverage by cold cloudiness and convection, greater upper-level divergence/upward motion, and greater moisture throughout a large depth of the troposphere are most important for tropical cyclogenesis and distinguishing DWs from NDWs.

In general, the evolution of waves valid over smaller longitude bands is similar to that over the full analysis domain. However, some differences were observed over certain longitude bands. In particular, waves that developed a tropical cyclone over the Caribbean region (i.e., Caribbean DWs) showed some indication of an increase in convective intensity with time, at least in some wave phases. In addition, large-scale, low-level convergence increased with time for Caribbean DWs, whereas DWs valid over the full analysis domain showed virtually no increase in low-level convergence in any wave phase.

Leppert et al. (2013) examined several indicators of convective coverage and intensity to determine which parameters provided the greatest distinction between DWs and NDWs over various fixed longitude bands (i.e., Eulerian approach). This study examined the same
parameters using a Lagrangian approach. In general, the results of both approaches agree with one another. For example, both approaches show that the greatest difference between DWs and NDWs is observed for the fractional coverage by IR brightness temperatures $\leq 240$ and $\leq 210$ K, while the indicators of convective intensity (i.e., lightning flash rates, mean cold PCTs, and mean convective reflectivity profiles) provide relatively few statistically significant differences between DWs and NDWs. However, results of this study indicate that on or near D0, lightning flash rates (PCTs) are significantly less (greater) for DWs, suggesting less intense convection during tropical cyclogenesis. Thus, the results of both this study and Leppert et al. (2013) suggest that the coverage by cold cloudiness/convection is generally more important than convective intensity for tropical cyclogenesis, although convective intensity may also be somewhat important for tropical cyclogenesis over the Caribbean or east Pacific.

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