An Examination of the Thermodynamic Impacts of Western North Pacific Tropical Cyclones on Their Tropical Tropospheric Environment

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ABSTRACT

The present study examines the tropospheric thermodynamic anomalies induced by western North Pacific tropical cyclone (TC) passage using storm-relative composites. Negative moist static energy (MSE) anomalies containing embedded westward-propagating anomalies generally occur only following larger TCs for two months following TC passage in a region extending from the domain center to ~3000 km to its west. Larger TCs force negative MSE anomalies likely because of feedbacks from stronger, broader TC-induced negative sea surface temperature (SST) anomalies and the excitation of TC-induced Rossby waves to the southeast of the TC. The negative MSE anomalies are composed of lower- and midtropospheric negative latent energy anomalies with smaller contributions from boundary layer and upper-tropospheric negative sensible heat anomalies. The lower- and midtropospheric negative MSE anomalies are forced by the TC, whereas the upper-tropospheric negative MSE anomalies are forced by the Madden–Julian oscillation. Vertically integrated MSE budgets at the domain center reveal negative MSE tendencies that are primarily forced by surface latent heat flux anomalies resulting from the TC-induced negative SST anomalies. Smaller negative MSE tendencies are due to 1) zonal and meridional advection of MSE anomalies by the Rossby waves and 2) enhanced top-of-the-atmosphere longwave radiative flux anomalies potentially associated with a reduction in the greenhouse gas effect of water vapor. The budget analysis in the west region is generally similar except that all terms are comparable in magnitude and relatively weaker. These results conservatively suggest that larger TCs can anomalously cool and dry their synoptic-scale environment for ~40 days following TC passage.

1. Introduction

Tropical cyclones (TCs) can impact both their oceanic and atmospheric environment on spatiotemporal scales that extend well beyond the TC. In the ocean, entrainment mixing, upwelling, and, to a lesser extent, surface latent and sensible heat fluxes by the TC yield a cooling of sea surface temperatures (SSTs) and the mixed layer along the TC track (e.g., Chang and Anthes 1978; Price 1981; Mao et al. 2000) lasting, on average, for several weeks following TC passage (Hazelworth 1968; Emanuel 2001; Hart et al. 2007; Price et al. 2008; Jansen et al. 2010; Dare and McBride 2011; Lloyd and Vecchi 2011; Mei and Pasquero 2012; Vincent et al. 2012a,b; Mei and Pasquero 2013; Vincent et al. 2013; Jullien et al. 2014). Entrainment mixing by the TC also yields positive temperature anomalies in the thermocline (e.g., Price 1981; Brooks 1983; Dickey et al. 1998; Zedler et al. 2002), which can last from several weeks to a year following TC passage (Emanuel 2001; Pasquero and Emanuel 2008; Jansen et al. 2010; Jullien et al. 2012; Mei et al. 2013; Vincent et al. 2013). These results suggest that TCs are able to significantly destratify their oceanic environment from between several weeks to a year following TC passage.

In contrast to the oceanic thermodynamic impacts of TCs, comparatively little is known about the potential atmospheric impacts of TCs. Previous studies have
suggested that TCs may be able to cool and dry their tropospheric environment for several weeks following TC passage (Sobel and Camargo 2005; Hart et al. 2007, 2008; Schenkel and Hart 2011; Jullien et al. 2014). However, none of these previous studies have comprehensively documented the four-dimensional structure of the tropospheric environmental anomalies following TC passage. Prior work has hypothesized that the negative temperature and moisture anomalies in the tropospheric environment following TC passage potentially result from three factors. First, a reduction in surface latent and sensible heat fluxes due to the negative SST anomalies induced by the TC may yield a cooling and drying of the lower troposphere (Sobel and Camargo 2005; Hart et al. 2007; Schenkel and Hart 2011; Jullien et al. 2014). Second, the secondary circulation of the TC may yield a drying of the troposphere due to subsidence outside the moist TC core and the anomalously high precipitation efficiency of the TC relative to isolated tropical convection (e.g., Palmén and Newton 1969; Frank 1977, 1982; Emanuel 1986, 1988, 2008; Houze 2010; Schenkel and Hart 2015). The subsidence and high precipitation efficiency of the TC secondary circulation may also suppress convection and reduce the associated latent heat release, yielding a cooling of the troposphere. Third, negative tropospheric moisture anomalies may indirectly yield tropospheric cooling owing to a reduction in the greenhouse gas effect of water vapor (Emanuel 2008).

In light of the uncertainties in the tropospheric thermodynamic anomalies following TC passage and the processes responsible for their generation, the present study seeks to address these issues using composites of several hundred western North Pacific (WPAC) TCs. Specifically, four-dimensional storm-relative composited data are used to quantify the spatiotemporal scales of the tropospheric thermodynamic anomalies following WPAC TC passage for comparison with prior work. Composited vertically integrated moist static energy (MSE) budgets are used to determine the processes and phenomena responsible for the tropospheric thermodynamic anomalies following WPAC TC passage. This study serves as a companion study to Schenkel and Hart (2015), which examined the large-scale environmental moisture impacts of TCs during TC passage.

The remainder of this study is divided into three parts. Section 2 describes the data and methodology. Section 3 quantifies the spatiotemporal scales of the MSE anomalies, examines the phenomena responsible for the MSE anomalies, and attributes the MSE anomalies to processes using MSE budgets. Section 4 summarizes the results and discusses future research.

2. Data and methods

a. Data

The response of the tropical tropospheric thermodynamic environment to WPAC TC passage is studied using storm-relative composites of reanalysis data. All 6-h Joint Typhoon Warning Center (JTWC) best-track (Chu et al. 2002) WPAC TCs over the ocean at or equatorward of 20°N from 1982 through 2009 are chosen for analysis. An upper-latitude threshold of 20°N is used to minimize the inclusion of subtropical and midlatitude features that intrude into the tropics. WPAC TCs are chosen for examination because these TCs are, on average, the largest, most intense, and most frequent TCs relative to all other basins (Gray 1968; Emanuel et al. 2004; Frank and Young 2007; Chavas and Emanuel 2010; Tory and Frank 2010; McTaggart-Cowan et al. 2013; Knaff et al. 2014), possibly suggesting that WPAC TCs have the greatest spatial and temporal impact on their tropospheric environment. The large size of WPAC TCs may also suggest that WPAC TCs and their environmental impacts are better resolved within reanalyses (Schenkel and Hart 2012). In this study, TC will refer to WPAC TCs unless noted otherwise.

The present study represents the atmosphere using the 0.5° × 0.5° 6-h National Centers for Environmental Prediction (NCEP) Climate Forecast System Reanalysis (CFSR; Saha et al. 2010). The CFSR is chosen because of its strong depiction of TC intensity (Schenkel and Hart 2012), track (Schenkel and Hart 2012), and structure (Wood and Ritchie 2014). Moreover, the CFSR is the only coupled atmosphere–ocean reanalysis (Saha et al. 2010) at the time of writing, potentially allowing for a more realistic representation of the effects of TC-induced negative SST anomalies upon the troposphere.

b. Methods

1) COMPOSITING

The present study uses storm-relative composites of CFSR data to examine the response of the tropical tropospheric thermodynamic environment during TC passage following the methodology of the companion study, Schenkel and Hart (2015). Please refer to Schenkel and Hart (2015) for a more detailed discussion of the compositing methodology used in the present study. At each reanalysis grid point, a 6-h 28-yr (1982–2009) climatological mean is computed for each variable and filtered in time using a daily 1–2–3–2–1 filter to reduce the contribution of phenomena with time scales of less than one week (Holland 1986). Next, composite TC-centered grids for each 6-h best-track TC data point are constructed from the reanalyses and binned according to
two different TC intensity groupings to examine the sensitivity of the environmental anomalies to best-track TC intensity: tropical storms (TS; 64 kt ≥ maximum 10-m wind speed ≥ 34 kt; \( N = 589 \) unique TCs) and typhoons (maximum 10-m wind speed ≥ 64 kt; \( N = 355 \) unique TCs). Table 1 contains the mean location, intensity, translation speed, date of occurrence, and number of cases for each TC intensity category. The TC-centered grids are fixed in space upon each best-track TC at 6-h intervals for 60 days prior through 60 days after TC passage (481 total grids) to allow for the examination of how the environment responds before, during, and after TC passage. In this study, any reference to TC passage refers to TC passage at the domain center unless explicitly stated otherwise.

The composite grid consists of 8000 km in the zonal direction, 7000 km in the meridional direction, and 38 vertical levels. Consistent with prior work (Schenkel and Hart 2011, 2015), the composite grid has a uniform horizontal grid spacing of 42 km, which is chosen to retain those features that are resolved in the subtropics and tropics within the reanalysis. The vertical grid spacing of the composite grid is 25 hPa from the surface to 100 hPa.

2) STATISTICAL SIGNIFICANCE

In the present study, a 10 000-sample bootstrap approach is used to determine whether the composited anomalies are statistically significantly different from zero at the 95% confidence interval using a two-tailed test consistent with prior work (e.g., Roundy and Frank 2004; Frank and Roundy 2006; Ventrice et al. 2011, 2012a,b; Ventrice and Thornicroft 2013; Schenkel and Hart 2015). Specifically, a new distribution of anomalies is constructed for a given variable at each grid point by randomly selecting anomalies (with replacement) for times in which a best-track TC is present. The sample size of each new distribution is conservatively defined according to the number of distinctly named TCs within the original sample rather than the number of 6-h best-track TC data points given the interdependence of 6-h best-track TC data points for a given TC. Specifically, any time a given TC is randomly selected, all of its 6-h data points are used to construct a new distribution while only counting once toward the sample size variable. A mean is then calculated for the new distribution of anomalies with the process being repeated until there are 10 000 means. The distribution of 10 000 means is then used to compute the 95% confidence interval of the mean using a two-tailed test to determine whether the anomalies are statistically significantly different from zero. It is important to note that the statistical-significance criterion does not discriminate between anomalies induced by the TC versus other phenomena [e.g., Madden–Julian oscillation (MJO) or convectively coupled equatorial waves], likely causing these phenomena to alter the statistical significance of the TC-induced anomalies. However, the alteration of the statistical significance of the TC-induced anomalies by other phenomena is believed to be minimal as suggested by the filtering of anomalies (see section 3).

3) MSE ANOMALIES AND BUDGETS

In the present study, MSE anomalies are used to evaluate how TCs thermodynamically impact their tropospheric environment:

\[
\text{MSE} = c_p T + Lq + g z, \tag{1}
\]

where \( c_p \) is the specific heat capacity of air at constant pressure, \( T \) is the temperature, \( L \) is the latent heat of vaporization of water, \( q \) is the specific humidity, \( g \) is the gravitational acceleration, and \( z \) is the geopotential height.

Vertically integrated MSE anomalies from the surface to 100 hPa are used to evaluate the tropospheric response, similar to prior work (Back and Bretherton 2006; Maloney 2009; Kiranmayi and Maloney 2011; Sobel et al. 2014).

Vertically integrated MSE budget anomalies, as given in Neelin and Held (1987), are calculated from the composites to identify the processes responsible for generating tropospheric thermodynamic anomalies:

\[
\frac{1}{g} \int_{p_t}^{p_0} \left( \frac{\partial \text{MSE}}{\partial t} \right) dp = -\frac{1}{g} \int_{p_0}^{p_t} (\mathbf{v} \cdot \nabla \text{MSE}) dp + \int_{p_0}^{p_t} \omega \frac{\partial \text{MSE}}{\partial p} dp + \text{LH}^t + \text{SH}^t + \text{LW}^t + \text{SW}^t, \tag{2}
\]

where \( p_t \) is the surface pressure, \( p_0 \) is the tropopause (100 hPa), \( \mathbf{v} \) is the horizontal velocity vector, \( \omega \) is the vertical pressure velocity, \( \text{LH}^t \) is the surface latent heat flux anomaly, \( \text{SH}^t \) is the surface sensible heat flux anomaly, \( \text{LW}^t \) is the vertically integrated longwave heating rate.

**TABLE 1.** TC mean best-track latitude, longitude, translation speed (m s\(^{-1}\)), maximum 10-m wind speed (kt; 1 kt ~ 0.51 m s\(^{-1}\)), Julian day of occurrence, number of unique TCs, and number of 6-h data points for TS and typhoons in the present study. The value to the right of the mean after the plus/minus sign is the standard error of the mean.

<table>
<thead>
<tr>
<th></th>
<th>TS</th>
<th>Typhoons</th>
</tr>
</thead>
<tbody>
<tr>
<td>Latitude (°N)</td>
<td>14.0 ± 0.2</td>
<td>15.5 ± 0.2</td>
</tr>
<tr>
<td>Longitude (°E)</td>
<td>133.4 ± 0.7</td>
<td>133.4 ± 0.8</td>
</tr>
<tr>
<td>TC translation speed</td>
<td>5.0 ± 0.1</td>
<td>5.0 ± 0.1</td>
</tr>
<tr>
<td>Max 10-m wind speed</td>
<td>45.2 ± 0.4</td>
<td>94.2 ± 1.3</td>
</tr>
<tr>
<td>Julian day</td>
<td>239.3 ± 3.5</td>
<td>253.1 ± 4.0</td>
</tr>
<tr>
<td>No. of unique TCs</td>
<td>589</td>
<td>355</td>
</tr>
<tr>
<td>No. of 6-h data points</td>
<td>4642</td>
<td>3946</td>
</tr>
</tbody>
</table>
anomaly, and $\text{SW}'$ is the vertically integrated shortwave heating rate anomaly. The term on the left-hand side of the budget equation represents the Eulerian time tendency of vertically integrated MSE anomalies. The first two terms on the right-hand side of the equation are the horizontal advection of MSE anomalies and vertical advection of MSE anomalies. The surface sensible and latent heat flux anomalies and longwave and shortwave heating rate anomalies are output by the reanalysis, while the remaining three terms are computed from the composite grid. To simplify the analysis, the surface latent and sensible heat flux anomalies and shortwave and longwave heating rate anomalies are initially grouped into a single term known as the “flux anomaly.”

The variables composing each budget term are separated into their climatological mean and deviation from the climatological mean to provide further insight into the composited anomalous MSE tendencies as exemplified below by the vertical advection of MSE anomalies:

$$-rac{1}{g} \int_{p_i}^{p_0} \left( \frac{\partial \text{MSE}}{\partial p} \right) \, dp \approx -rac{1}{g} \int_{p_i}^{p_0} \omega \frac{\partial \text{MSE}}{\partial p} \, dp$$

$$- \frac{1}{g} \int_{p_i}^{p_0} \frac{\partial \text{MSE}'}{\partial p} \, dp$$

$$- \frac{1}{g} \int_{p_i}^{p_0} \omega \frac{\partial \text{MSE}'}{\partial p} \, dp. \tag{3}$$

The overbar quantities represent the aforementioned composited 6-h climatological mean of a given variable, while the primed quantities represent the composited 6-h deviation from the climatological mean of a variable. The separation of the variables into their climatological mean and deviation from the climatological mean is used to determine how MSE anomalies are forced by interactions between the TC and its environment. It should be noted that while the TC is the dominant feature in the deviation from the climatological mean within the storm-relative composites, contributions from other phenomena are also present. In the present study, the terms containing products of means are ignored because they primarily describe time scales much longer than the tropospheric thermodynamic anomalies induced by the TC, as also suggested in prior work (e.g., Sobel and Camargo 2005; Hart et al. 2007, 2008; Schenkel and Hart 2011; Jullien et al. 2014).

To isolate the impacts of TCs from the MJO and equatorial waves, zonal space–time spectral filtering of anomalies for a given variable is used following the methodology of Wheeler and Kiladis (1999). A band filter similar to Wheeler and Kiladis (1999) is used for the MJO; Kiladis et al. (2009) is used for Kelvin waves, equatorial Rossby (ER) waves, and mixed Rossby–gravity (MRG) waves; and Frank and Roundy (2006) is used for tropical depression (TD)-type waves. However, it is important to note that these wavenumber–frequency filter bounds are determined from top-of-the-atmosphere (TOA) upwelling longwave radiative fluxes and may not necessarily be representative of vertically integrated MSE anomalies used in the present study.

c. Justification of the use of reanalyses for studying environmental impacts of TCs

Prior to discussing the results, it is important to briefly justify the use of reanalyses to study TCs in the present study. Prior work has shown that reanalyses may introduce substantial uncertainty into the present study primarily owing to two deficiencies: 1) representation of MSE and 2) representation of TCs. Specifically, deficiencies in reanalysis MSE budgets are primarily due to the non-conservation of energy (e.g., Chiodo and Haimberger 2010; Trenberth et al. 2011) that results from the assimilation of new data every 6 h (Bloom et al. 1996) and the substantial differences that exist between observed and reanalysis radiative fluxes, surface fluxes, precipitation, and moisture (Onogi et al. 2007; Bosilovich et al. 2008; Chiodo and Haimberger 2010; Berrisford et al. 2011; Bosilovich et al. 2011; Trenberth et al. 2011). Deficiencies in reanalysis TC representation primarily result from muted reanalysis TC intensity and structure (Hatsushika et al. 2006; Manning and Hart 2007; Onogi et al. 2007; Schenkel and Hart 2012, 2015; Wood and Ritchie 2014) relative to the best track because of the coarse grid spacing and conservative microphysical and convective parameterizations used in reanalyses.

In spite of these caveats associated with reanalyses, there are three reasons why there is reasonable confidence in the use of reanalyses in the present study. First, reanalyses have been used in prior work to study the energetics of phenomena including the MJO (e.g., Kiladis et al. 2005; Kiranmayi and Maloney 2011; Sobel et al. 2014) and TCs (e.g., Sobel and Camargo 2005; Hart et al. 2007, 2008; Schenkel and Hart 2011, 2015). Second, the present study is also dependent on the representation of the large-scale TC structure (i.e., warm-core structure on the meso-α to synoptic scale), which is relatively well resolved (Schenkel and Hart 2015; Wood and Ritchie 2014). Last, any substantial errors in TC representation are nudged back to observations by the reanalysis data assimilation (Thorne and Vose 2010). For a more comprehensive justification of the use of reanalyses in the present study, please see Schenkel and Hart (2015).

3. Results

Sections 3a–e examine the horizontal, temporal, and vertical scales of the tropospheric MSE anomalies before
and after TC passage, the energy components of the MSE anomalies, and the atmospheric phenomena responsible for these anomalies. Sections 3f and 3g examine the spatiotemporal structure of vertically integrated MSE budgets to attribute the negative MSE anomalies to processes.

a. Horizontal and temporal structure of MSE anomalies

The horizontal and temporal structure of the tropospheric thermodynamic TC environment before, during, and after TC passage is depicted using time–longitude plots of vertically integrated MSE anomalies along the latitude of TC passage (Fig. 1a) and typhoon passage (Fig. 1b). The green lines labeled TC and W1–W6 denote the TC and TC-induced Rossby waves, respectively, and are referenced in the text. Anomalies are shown only if they are statistically significantly different from zero at the 95% confidence interval.
Schenkel and Hart 2011). However, the tropospheric environmental anomalies before TC passage in these previous studies are small, with several studies neglecting to calculate their significance (Hart et al. 2007, 2008; Schenkel and Hart 2011).

During TC passage, the eastern half of the domain is characterized by large-scale positive MSE anomalies from which the TC emerges (green line labeled with TC) with a large-scale region of negative MSE anomalies to the west of the TC (Sobel and Camargo 2005; Schenkel and Hart 2015). Prior work has shown that the dipole in MSE anomalies during TC passage is due to the MJO, ER wave, and TCs (Schenkel and Hart 2015). The MSE anomalies during TC passage do not directly influence the MSE anomalies beyond several days after TC passage and are not discussed further.

The tropospheric environment after TC passage is characterized by negative MSE anomalies generally extending from the domain center to ~3000 km to the west for two months following TC passage, consistent with prior work (Schenkel and Hart 2011). The timing of negative MSE anomaly onset relative to TC passage suggests that the negative MSE anomalies are, at least, partially forced by the TC. The horizontal area of the negative MSE anomalies gradually shrinks with increasing time from TC passage. Both the TS (Fig. 1a) and typhoon composite (Fig. 1b) contain westward-propagating negative MSE anomalies that are maximized to the west of the domain center (green lines labeled with W2–W6) following TC passage consistent with previous work (Hart et al. 2007, 2008; Schenkel and Hart 2011), although the identity of the westward-propagating anomalies are undefined in these prior studies.

It is also instructive to examine SST anomalies since the negative SST anomalies induced by the TC and the resulting reduction in sea surface fluxes are hypothesized to play a strong role in forcing negative tropospheric MSE anomalies following TC passage (Sobel and Camargo 2005; Hart et al. 2007, 2008; Schenkel and Hart 2011; Julien et al. 2014). Specifically, time–longitude plots of SST anomalies along the latitude of TS (Fig. 2a) and typhoon (Fig. 2b) passage are shown. Negative SST anomalies are present for over two months following TS and typhoon passage, consistent with prior work (Dare and McBride 2011; Lloyd and Vecchi 2011; Mei and Pasquero 2012; Vincent et al. 2012a,b; Mei and Pasquero 2013; Vincent et al. 2013; Julien et al. 2014). The horizontal scale of slightly less than ~3500 km for the negative SST anomalies is comparable to the negative MSE anomalies (Figs. 1a,b) and is consistent with prior work (Mei and Pasquero 2013). The negative SST anomalies are also generally centered to the east of the negative MSE anomalies (Fig. 1), suggesting that parcels are anomalously cooled and dried because of the resulting reduction in sea surface fluxes, as they are advected westward by climatological lower-tropospheric easterlies [see section 3f(1)]. A comparison of Figs. 2a,b reveals that negative SST anomalies are generally stronger for typhoons relative to TS.

These results suggest tropospheric MSE anomalies are initially near climatology prior to TC passage before becoming negative over broad spatial and temporal scales following TC passage. The TC-induced negative SST anomalies are likely one source of forcing for negative MSE anomalies, but the smoothly varying SSTs do not explain the variability in the negative MSE anomalies following TC passage. Section 3b focuses on the variability of the negative MSE anomalies following typhoon passage.

b. Variability in the horizontal structure of tropospheric MSE anomalies

Close examination of the time–longitude plots of vertically integrated MSE anomalies for typhoons (Fig. 1b) reveals pulsations in the negative MSE anomalies every ~ (6−18) days throughout the 60 days following TC passage, consistent with prior work (Hart et al. 2007, 2008; Schenkel and Hart 2011). These pulsations in negative MSE anomalies originate to the east of the domain center and propagate westward at ~ (5−6) m s⁻¹. The increased frequency and magnitude of the westward-propagating anomalies following TC passage suggest that these negative MSE anomalies are triggered by TC passage.

Additional insight into the westward-propagating anomalies is provided by examining time–longitude plots of 850-hPa meridional wind anomalies along the latitude of TS (Fig. 3a) and typhoon (Fig. 3b) passage. Westward-propagating anomalies are generally collocated with lower-tropospheric northerly wind anomalies (Fig. 3b) that weaken with height and extend into the upper troposphere (not shown). Additionally, the first northerly wind anomaly following typhoon passage (W2 in Fig. 3b) begins to the east of the domain center with each successive anomaly moving farther east, which suggests that the northerly wind anomalies have an eastward group velocity (black arrow in Fig. 3b).

Zonal space–time spectral filtering of the MSE anomalies reveals that the westward-propagating anomalies project, at times, onto the ER wave band filter as shown by the black contours in the time–longitude plot of MSE anomalies (Fig. 1). The presence of ER waves after TC passage within the composite is not surprising given that ER waves occur during 25%−55% of WPAC TC genesis events (Frank and Roundy 2006; Schreck et al. 2011, 2012). However, zonal space–time spectral filtering generally explains only, at most, less than 50% of the negative MSE
anomalies, suggesting that alternative phenomena are also partially responsible for the westward-propagating anomalies. It remains plausible that the ER waves may partially be an artifact of filtering MSE anomalies containing a TC, which strongly projects onto the ER wave band filter (Schreck et al. 2011; Aiyyer et al. 2012; Schreck et al. 2012).

Further insight into the horizontal structure of MSE anomalies is obtained by examining plan view plots before (day $+10.5$), during (day $+13.5$), and after (day $+16.5$) the passage of westward-propagating anomalies at the domain center. Specifically, plan view plots of vertically integrated total MSE and MSE anomalies at day $+10.5$ for TS (Fig. 4a) and typhoons (Fig. 4b), at day $+13.5$ for TS (Fig. 4c) and typhoons (Fig. 4d), and at day $+16.5$ for TS (Fig. 4e) and typhoons (Fig. 4f) are shown. The regions labeled N1 and W1–W2 in Fig. 4 denote a preexisting area of negative MSE anomalies and westward-propagating MSE anomalies, respectively. The negative MSE anomalies associated with the westward-propagating anomalies (W2 in Fig. 4) strengthen in magnitude and expand in horizontal scale as they propagate west-northwestward through and to the west of the domain center in both the TS and typhoon composite. Upon reaching the domain center, these westward-propagating anomalies encompass a region of thousands of square kilometers (Figs. 4c,d). These results suggest that westward-propagating anomalies are
not merely a smearing of increasingly out-of-phase pulses in space and time in the composite but instead the passage of a coherent wave at a specific wavelength and frequency in the majority of typhoons constituting the composite. In both the TS and typhoon composite, the westward-propagating anomalies merge with a preexisting area of negative MSE anomalies (N1 in Figs. 4a–c), yielding a larger and stronger area of negative MSE anomalies. In the typhoon composites, the preexisting region of negative MSE anomalies develops at the domain center immediately following TC passage, suggesting that these anomalies are forced by the TC. Following the merging of the preexisting MSE anomalies and westward-propagating MSE anomalies, the westward-propagating anomalies weaken and dissipate. These results have documented the presence of local minima in the vertically integrated MSE anomalies that propagate west-northwestward and occasionally project unto the ER wave band filter. While the growth in the magnitude and area of these westward-propagating anomalies as they move through the domain center suggests that these anomalies may be modulated by the TC, uncertainties remain regarding the identity of these westward-propagating MSE anomalies. Section 3c will

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**FIG. 3.** As in Fig. 1, but for 850-hPa meridional wind anomalies (m s$^{-1}$). The black arrow denotes the approximate eastward group velocity of the meridional wind anomalies.
explore the potential phenomena responsible for the westward-propagating anomalies.

**c. Source of the westward-propagating negative MSE anomalies**

Before continuing, it is necessary to introduce four regions to be analyzed throughout this study (labeled in Fig. 4a): 1) domain center, 2) west region, 3) south region, and 4) southwest region. The domain center is chosen to study how the environment responds to the TC directly passing through it. The west region is chosen because it corresponds to the location of the strongest negative MSE anomalies. The south and southwest regions are chosen as regions that are characteristic of TC
environment while not being substantially impacted by the TC. Each of the four regions is 500 km by 500 km, approximately corresponding to the box size used in Hart et al. (2007).

To examine to what degree subsequent TC activity is responsible for the westward-propagating anomalies, a time series of TC frequency is constructed for TCs passing within a 1700-km radius of the domain center and west region between 15% and 30% of the time in the weeks after TC passage to potentially impact these regions with pulsations in TC frequency occurring in the weeks after typhoon passage at a comparable frequency to the westward-propagating anomalies (~6–18 days; black vertical lines labeled with W2–W6 in Fig. 5). However, the westward-propagating anomalies are coincident with minima in TC frequency at the domain center while exhibiting a weaker relationship at the west region. The comparatively small differences in TC activity before and after TC passage also suggest that changes in TC activity are not responsible for the negative MSE anomalies. These results suggest that subsequent TC activity is not responsible for generating the negative MSE anomalies or westward-propagating anomalies.

An additional potential source of the westward-propagating anomalies may be Rossby wave dispersion induced by the TC to its south and east (e.g., Davidson and Hendon 1989; Shapiro and Ooyama 1990; Ritchie and Holland 1999; Li and Fu 2006; Fu et al. 2007; Ge et al. 2007; Krouse et al. 2008; Krouse and Sobel 2010). Prior work has suggested that Rossby waves are excited to the south and east of the TC along the environmental planetary vorticity gradient when the TC is propagating more quickly westward than the lower-tropospheric flow to its south and east (Krouse et al. 2008; Krouse and Sobel 2010). TCs can propagate more quickly westward than the lower-tropospheric flow under two sets of environmental conditions (Ge et al. 2007; Krouse et al. 2008): 1) lower-tropospheric cyclonic zonal wind shear and, to a lesser extent and 2) easterly vertical wind shear. With regard to the first condition, the dispersion of Rossby waves to the south and east of the TC (e.g., Davidson and Hendon 1989; Shapiro and Ooyama 1990; Ritchie and Holland 1999; Li and Fu 2006; Fu et al. 2007; Ge et al. 2007; Krouse et al. 2008; Krouse and Sobel 2010) suggests that the cyclonic zonal shear yields an environmental steering flow that is more strongly westward for the TC relative to the Rossby waves given that the TC is generally farther north and propagating more quickly northward. For the second condition, the environmental steering flow of the TC (e.g., Dong and Neumann 1986; Velden and Leslie 1991) extends over a greater depth of the troposphere than the Rossby waves (e.g., Li and Fu 2006; Fu et al. 2007; Ge et al. 2007) because of the greater

![Figure 5](http://journals.ametsoc.org/doi/pdf/10.1175/JCLI-D-14-00780.1)
vertical depth of the TC. As a result, the environmental steering flow for the TC will be more strongly westward in an environment with easterly vertical wind shear. Easterly vertical wind shear also tends to support stronger Rossby wave dispersion by trapping Rossby wave energy in the lower troposphere (Wang and Xie 1996; Ge et al. 2007).

Given that the broad horizontal scale of the TC circulation (i.e., lower-tropospheric cyclonic circulation of TC can exceed 1700-km radius; Knaff et al. 2014) can potentially contaminate the calculation of environmental vertical and zonal wind shear, Krouse and Sobel (2010) carefully selected a simplified set of parameters calculated from reanalyses and averaged over the TC lifetime based upon prior idealized modeling studies (Ge et al. 2007; Krouse et al. 2008) to represent environmental vertical and zonal wind shear. Specifically, environmental vertical wind shear is calculated by area averaging 850–200-hPa vertical wind shear over a 2500 km × 2500 km box centered on the TC (Krouse and Sobel 2010). Environmental zonal wind shear is calculated by taking the difference between the zonal propagation speed of the TC and environmental zonal steering flow of the Rossby waves, with the latter parameter calculated by area averaging over a 2400 km × 1000 km box centered to the southeast of the TC (Krouse and Sobel 2010). The results of Krouse and Sobel (2010) suggest that TC-induced Rossby waves are favored for environmental cyclonic zonal wind shear and easterly or weak westerly vertical wind shear.

Using the criteria of prior work (Krouse and Sobel 2010), the present study constructs composites of TCs that are present in environments that are favorable and unfavorable for TC-induced Rossby waves. Specifically, zonal and vertical wind shear, calculated using the criteria of Krouse and Sobel (2010), must be greater than or equal to 2 and −5 m s⁻¹, respectively, for the environment to be considered favorable for TC-induced Rossby waves. Table 2 contains the mean location, intensity, translation speed, date of occurrence, and number of cases for each TC subset in the favorable and unfavorable TC-induced Rossby wave dispersion composites, with substantial differences occurring only for sample size. It is important to note that ~34% of TCs constitute the favorable composite, which is comparable to the ~41% of WPAC TCs identified as triggering Rossby waves (Li and Fu 2006; Fu et al. 2007), thus providing confidence that the thresholds of Krouse and Sobel (2010) have correctly separated those cases with and without TC-induced Rossby waves.

A time–longitude plot of vertically integrated MSE anomalies at the latitude of typhoon passage for typhoons with unfavorable (Fig. 6a) and favorable (Fig. 6b) conditions is shown. Similar to Fig. 1, the green lines denote the TC and the westward-propagating MSE anomalies (W1–W7). Figure 6b reveals that the favorable composite contains stronger negative MSE anomalies and stronger pulsations (W2–W7 in Fig. 6b) compared to the original typhoon composite (Fig. 1b). In contrast, the unfavorable composite (Fig. 6a) does not have any MSE anomalies following TC passage demonstrating the variability in the environmental response following TC passage.

Further insight is provided by examining time–longitude plots of SST anomalies at the latitude of typhoon passage for the unfavorable (Fig. 7a) and favorable (Fig. 7b) composites. The favorable composite exhibits negative SST anomalies that are both larger in magnitude and horizontal extent than the unfavorable composite, with the resulting surface latent heat flux anomalies (not shown) possibly explaining the differences in negative MSE anomalies between the two composites. The comparable TC intensity, location, date of occurrence, and TC translation speed in the favorable and unfavorable composites (Table 2) suggest that the differences in negative SST anomalies between the composites are likely due to TC size differences. It is important to note that prior work has shown that larger TCs disperse Rossby waves more strongly, consistent with the differences in the composites (Carr and Elsberry 1995).

Lastly, time–longitude plots of 850-hPa meridional wind anomalies at the latitude of typhoon passage for the unfavorable (Fig. 8a) and favorable (Fig. 8b) composites are shown. As expected from Fig. 6b, the favorable composite exhibits stronger westward-propagating northerly wind anomalies (W2–W7 in Fig. 8b). Moreover, the eastward group velocity of the meridional wind anomalies, which is consistent with TC-induced Rossby waves (e.g., Davidson and Hendon 1989; Shapiro and Ooyama 1990; Ritchie and Holland 1999; Li and Fu 2006; Fu et al. 2007; Krouse et al. 2008; Krouse and Sobel 2010), is more clearly observed in the favorable composite for typhoons (black line in Fig. 8b) compared to the original composite. In contrast, the unfavorable composite does not contain any coherent westward-propagating meridional wind anomalies (Fig. 8a).
These results suggest that the source of the westward-propagating anomalies is likely Rossby wave dispersion induced by a subset of relatively large typhoons. For the remainder of the present study, the westward-propagating MSE anomalies will be referred to as Rossby waves. More importantly, these results suggest that larger TCs generally can excite negative MSE anomalies following TC passage owing to their ability to trigger Rossby waves and a broad region of negative SST anomalies, whereas smaller TCs generally do not impact their tropospheric environment. Section 3d examines the vertical structure of the negative MSE anomalies and its components as a preliminary step toward determining the contribution of the TC-induced SST anomalies and Rossby waves to the negative MSE anomalies.

Fig. 6. Time–longitude plot of vertically integrated MSE anomalies (10^6 J m^-2; shaded) for typhoons in a large-scale environment that is (a) unfavorable and (b) favorable for TC-induced Rossby waves meridionally averaged between 250 km to the south and 250 km to the north of the latitude of TC passage. The criteria used for determining whether the large-scale environmental conditions are unfavorable or favorable for TC passage are from prior work (Krouse et al. 2008; Krouse and Sobel 2010). The green lines denote the TC and TC-induced Rossby waves and are referenced in the text. Anomalies are shown only if they are statistically significantly different from zero at the 95% confidence interval.
d. Vertical structure of MSE anomalies

The vertical and temporal structure of the thermodynamic anomalies are shown in time–height plots of area-averaged MSE anomalies at the domain center for TS (Fig. 9a) and typhoons (Fig. 9b), latent energy anomalies for TS (Fig. 9c) and typhoons (Fig. 9d), and sensible heat anomalies for TS (Fig. 9e) and typhoons (Fig. 9f). Similar time–height plots of area-averaged MSE anomalies at the west region for TS (Fig. 10a) and typhoons (Fig. 10b), latent energy anomalies for TS (Fig. 10c) and typhoons (Fig. 10d), and sensible heat anomalies for TS (Fig. 10e) and typhoons (Fig. 10f) are also shown. The black lines labeled TC and W1–W5 denote the TC and TC-induced Rossby waves, respectively, in Figs. 9 and 10, while L, M, and U denote salient regions of lower-tropospheric, midtropospheric, and upper-tropospheric anomalies, respectively. Potential energy anomalies are not shown because they are an order of magnitude less than the latent energy and sensible heat anomalies. Figures 9a,b and 10a,b reveal negative MSE anomalies throughout the troposphere at the domain center and west region following both TS and typhoon passage. These negative MSE anomalies are largely composed of negative latent energy anomalies in the lower and midtroposphere (labeled with L and M in Figs. 9c,d and 10c,d) and,
to a lesser extent, negative sensible heat anomalies in the boundary layer and upper troposphere (labeled with L and U in Figs. 9c,f and 10c,f), consistent with prior work (Sobel and Camargo 2005; Hart et al. 2007, 2008; Schenkel and Hart 2011). The negative MSE anomalies are stronger throughout the troposphere for typhoons than for TS at the domain center and west region, also consistent with prior work (Hart et al. 2007, 2008). The Rossby waves (black lines in Figs. 9 and 10) are also likely responsible for the strong pulsations in negative MSE and latent energy anomalies in the lower and midtroposphere at the domain center and west regions for TS and typhoons, resembling the anomalies in prior work (Hart et al. 2007, 2008; Schenkel and Hart 2011). It is important to note that the negative MSE anomalies at the domain center only consistently occur in the boundary layer and the upper troposphere at the domain center (Figs. 9a,b) and at the west region for TS (Fig. 10a) during the first several weeks following TC passage.

At the domain center, the strongest negative MSE anomalies are concentrated in the boundary layer immediately following TC passage (L in Figs. 9a,b), similar to prior work (Schenkel and Hart 2011). The boundary layer MSE anomalies are composed of negative latent heat.

**FIG. 8.** As in Fig. 6, but for 850-hPa meridional wind anomalies (m s$^{-1}$). The black arrow denotes the approximate eastward group velocity of the Rossby waves.
Fig. 9. Time–height plots at the domain center of area-averaged MSE anomalies, latent energy anomalies, and sensible heat anomalies ($10^2$ J kg$^{-1}$; shaded) for (a),(c),(e) TS and (b),(d),(f) typhoons, respectively. Anomalies are shown only if they are statistically significantly different from zero at the 95% confidence interval. The labels L, M, and U denote the approximate heights and onset times of salient lower-tropospheric, midtropospheric, and upper-tropospheric anomalies, respectively, and are referenced in the text. The vertical black lines labeled with TC and W1–W5 correspond to the TC and TC-induced Rossby waves and are also referenced in the text. Gray shading denotes areas that are below the surface.
energy anomalies that are strongest at the top of the boundary layer (L in Figs. 9c,d) and negative sensible heat anomalies that are strongest near the surface (L in Figs. 9e,f). This structure is consistent with the expected negative latent energy and sensible heat anomalies due to a reduction of sea surface fluxes induced by negative SST anomalies (Hashizume et al. 2002; Small et al. 2003; Koseki and Watanabe 2010), in this case by the TC. In...
contrast, the negative MSE anomalies at the west region following TC passage are strongest in the midtroposphere (M in Figs. 10a,b), likely owing to stronger impacts by the Rossby waves (black vertical lines labeled with W2–W5 in Fig. 10). The sea surface fluxes induced by the TC are also likely weaker in the west region as expected given the relatively weaker negative SST anomalies (Figs. 2a,b). However, the vertical structure of the boundary layer negative latent energy (L in Figs. 10c,d) and sensible heat (L in Figs. 10e,f) anomalies at the west region suggest that the anomalous sea surface fluxes induced by the TC are still important.

These results suggest that the TC-induced negative SST anomalies force the negative MSE anomalies in the boundary layer, which are manifested as negative latent energy anomalies and, to a lesser extent, sensible heat anomalies. Moreover, the Rossby waves act to periodically excite negative MSE anomalies primarily as negative latent energy anomalies in the midtroposphere. However, the source of the negative upper-tropospheric MSE and sensible heat anomalies within the first several weeks following TC passage remains unclear and is the focus of section 3e.

e. Influence of the MJO upon the composited MSE anomalies

To determine the source of the negative upper-tropospheric MSE and sensible heat anomalies, the vertical structure of MSE anomalies is examined for the south and southwest regions as they typify the tropical environment outside the TC. Specifically, time–height plots of area-averaged MSE anomalies for the south (Fig. 11a) and southwest regions (Fig. 11b), latent energy anomalies for the south (Fig. 11c) and southwest regions (Fig. 11d), and sensible heat anomalies for the south (Fig. 11e) and southwest regions (Fig. 11f) are constructed for the typhoon composite. An analysis of Fig. 11 reveals that the south and southwest regions exhibit upper-tropospheric negative sensible heat anomalies at a similar time, location, and magnitude to the upper-tropospheric negative sensible heat anomalies at the domain center and west region. These results suggest that these anomalies are forced by large-scale phenomena (e.g., MJO) rather than the TC since the anomalies extend 2000 km from the domain center to the south region, which is well beyond the mean documented size of TC-induced upper-tropospheric sensible heat anomalies (Frank 1977). In contrast, the absence of lower- and midtropospheric MSE anomalies after several weeks in the south and southwest region suggests that these anomalies are unique to the TC track and are likely partially forced by the TC.

Additional insight is provided by examining plan view plots of 300-hPa sensible heat anomalies before (6 days prior to typhoon passage; Fig. 12a), during (Fig. 12b), and after (12 days after typhoon passage; Fig. 12c) typhoon passage. A time–longitude plot of 300-hPa sensible heat anomalies at the latitude of strongest negative sensible heat anomalies is also presented in Fig. 12d. TC, MJO+, and MJO— in Fig. 12 denote the TC, the convectively active MJO, and the convectively suppressed MJO, respectively. Figure 12 reveals a transition from a dipole in sensible heat anomalies prior to TC passage (Figs. 12a,d) to large-scale positive sensible heat anomalies during TC passage (Figs. 12b,d), to large-scale negative sensible heat anomalies in the weeks following TC passage (Figs. 12c,d). Zonal space–time spectral filtering reveals that the negative upper-tropospheric sensible heat anomalies following TC passage project onto the MJO band filter (black contours and MJO— in Fig. 12d). To provide further context to MJO activity in the composites, the real-time multivariate MJO index (RMM; Wheeler and Hendon 2004) is used, revealing that ~45% of typhoons are associated with a convectively suppressed MJO (RMM phases 1–3 and 8 with amplitude greater than 1) in the WPAC 12 days after typhoon passage, consistent with prior work (Frank and Roundy 2006). However, the remaining 55% of typhoons occur during either weak-amplitude MJOs or a convectively active MJO (RMM phases 4–7 with amplitude greater than 1) 12 days after typhoon passage, likely explaining the weak amplitude of the MJO-filtered negative upper-tropospheric sensible heat anomalies compared to prior work (e.g., Kiladis et al. 2005; Zhang 2005). Stratifying composite cases by RMM phase (not shown) does not reveal an environmental response resembling the composites presented here, suggesting that factors intrinsic to the TC (e.g., TC intensity, Rossby wave dispersion) are more important in determining the environmental response. On a similar note, a substantial fraction (~50%) of the negative upper-tropospheric sensible heat anomalies remain unexplained by zonal space–time spectral filtering in the MJO band filter, potentially suggesting that the anomalies may also be forced by the TC.

These results suggest that the MJO is partially responsible for the negative upper-tropospheric sensible heat anomalies following TC passage, although a contribution from TCs cannot be ruled out. The TC and its Rossby waves along with the MJO have been shown to contribute to the negative MSE anomalies, although the processes responsible for these anomalies remain unidentified. With this in mind, section 3f examines the processes responsible for the negative MSE anomalies using vertically integrated MSE budgets.

f. Horizontal structure of MSE budget tendency anomalies

The analysis of the composited vertically integrated MSE budgets focuses on the processes responsible for
the negative MSE anomalies following TC passage. Plan view plots of MSE budget terms are presented at days $+10.5$ and $+13.5$ following TC passage to demonstrate how TCs impact their environment. Particular emphasis is placed on the MSE tendencies at the domain center and west region. Days $+10.5$ and $+13.5$ after TC passage are chosen for analysis because these days are characteristic of the environmental response to the
Rossby waves in the weeks following TC passage. In the interest of brevity, MSE budgets are presented only for typhoons, although the results are similar, but muted, for TS.

The MSE budget time tendency analysis begins with plan view plots before (day +10.5) and after (day +13.5) the propagation of the first Rossby wave at the domain center for the vertical advection of MSE anomalies at days +10.5 (Fig. 13a) and +13.5 (Fig. 13b), horizontal advection of MSE anomalies at days +10.5 (Fig. 13c) and +13.5 (Fig. 13d), and flux anomalies at days +10.5 (Fig. 13e) and +13.5 (Fig. 13f). W2, H1–H3, and F1

FIG. 12. Plan view of 300-hPa sensible heat anomalies (10^7 J kg⁻¹):(a)–(c) at 6 days before, during, and at 12 days after typhoon passage, respectively. (d) Time–longitude plot of 300-hPa sensible heat and MJO band-filtered sensible heat anomalies (10^7 J kg⁻¹) meridionally averaged between y = −2142 and −1638 km from the domain center [purple lines in (a)–(c)] for typhoons. Anomalies are shown only if they are statistically significantly different from zero at the 95% confidence interval. The labels TC, MJO+, and MJO− denote the TC location, convectively active MJO, and convectively suppressed MJO, respectively, and are also referenced in the text. The green boxes denote the location of the domain center and west region, while the red boxes denote the location of the south and southwest regions in (a)–(c).
FIG. 13. Plan view of vertical advection and horizontal advection of MSE and flux anomalies (W m$^{-2}$) at (a),(c),(e) day +10.5 and (b),(d),(f) day +13.5 after typhoon passage, respectively. The labels W2, H1–H3, and F1 denote the TC-induced Rossby wave, salient regions of horizontal advection of MSE anomalies, and salient regions of flux anomalies, respectively, and are referenced in the text. The green boxes denote the location of the domain center and west region, while the red boxes denote the location of the south and southwest regions.

denote the TC-induced Rossby wave, salient regions of horizontal advection of MSE anomalies, and salient regions of flux anomalies, respectively, in Fig. 13. Vertical advection of MSE anomalies is positive (Figs. 13a,b) in both the domain center and west region helping to restore the negative MSE anomalies back to climatology and are not discussed any further given the focus of the analysis on the processes responsible for the negative MSE anomalies. At the domain center, the flux anomalies are the dominant source of negative MSE anomalies (F1 in Figs. 13e,f),
while both the flux and horizontal advection of MSE anomalies generate negative MSE anomalies at the west region (H1 in Figs. 13c,d and F1 in Figs. 13e,f). Both the horizontal advection of MSE anomalies and, to a lesser extent, the flux anomalies are sensitive to the passage of the Rossby waves at the domain center and west region. Section 3f(1) provides further insight into the phenomena responsible for the horizontal advection of MSE anomalies by partitioning this term into its components, which is followed by a similar analysis for the flux anomalies.

1) PHYSICAL INTERPRETATION OF HORIZONTAL ADOVCTION OF MSE ANOMALIES

To begin, the horizontal advection of MSE anomalies is partitioned into a zonal component at days +10.5 (Fig. 14a) and +13.5 (Fig. 14b) and a meridional component at days +10.5 (Fig. 14c) and +13.5 (Fig. 14d). Insight into the zonal and meridional advection of MSE anomalies is provided by decomposing the variables within each term into their climatological mean and deviation from the climatological mean. The decomposition of the zonal and meridional advection of MSE yields two dominant terms: the advection of anomalous MSE by the climatological zonal wind at days +10.5 (Fig. 15a) and +13.5 (Fig. 15b) and the advection of climatological MSE by the anomalous meridional wind at days +10.5 (Fig. 15c) and +13.5 (Fig. 15d). W2, Z1–Z2, and M1 denote the TC-induced Rossby wave, salient regions of zonal advection of MSE anomalies, and salient regions of meridional advection of MSE anomalies, respectively, in Figs. 14 and 15. Plan view plots of 1000–500-hPa layer-averaged total and anomalous zonal wind at days +10.5 (Fig. 16a) and +13.5 (Fig. 16b) and total and anomalous meridional wind at days +10.5 (Fig. 16c) and +13.5 (Fig. 16d) are provided to help interpret the results for horizontal advection of MSE anomalies. The 1000–500-hPa layer is chosen because it represents the
layer of strongest MSE anomalies. W2 and V1 denote the TC-induced Rossby wave and salient regions of meridional wind anomalies, respectively, in Fig. 16.

The horizontal advection of MSE anomalies following TC passage can be characterized by two regimes: 1) without Rossby waves and 2) with Rossby waves. In the absence of Rossby waves, the horizontal advection of MSE anomalies at the domain center is largely dominated by positive zonal advection of MSE anomalies (Z2 in Fig. 14b). The zonal advection of MSE anomalies results from the downgradient advection of relatively higher MSE into the domain center (Fig. 4b) by the lower- and midtropospheric climatological easterly wind (Fig. 16b), as seen by Z2 in Fig. 15b. Horizontal advection of MSE anomalies at the west region, in the absence of Rossby waves, is negative in the second week following TC passage because of the zonal advection of MSE anomalies (Z1 in Fig. 14a). The zonal advection of MSE anomalies results from the upgradient advection of negative MSE anomalies at the domain center (N1 in Fig. 4b) by the lower- and midtropospheric climatological easterly wind (Fig. 16a), as seen by Z1 in Fig. 15a. During subsequent times, the anomalous zonal MSE gradient (Fig. 1b) is too weak to yield substantial zonal advection of MSE anomalies at the west region except during Rossby wave passage.

During Rossby wave passage, meridional advection of MSE anomalies (M1 in Figs. 14c,d) and, to a lesser extent, zonal advection of MSE anomalies (Z2 in Fig. 14a) by the Rossby waves (W2 in Fig. 14) modulate the horizontal advection of MSE anomalies at the domain center and west region. The meridional advection of MSE anomalies results from northerly wind anomalies
(V1 in Figs. 16c,d) associated with the Rossby waves imposed upon the climatological meridional MSE gradient (Figs. 4b,d,f), as seen by M1 in Figs. 15c,d. Zonal advection of MSE anomalies results from the advection of the negative MSE anomalies associated with the Rossby waves (W2 in Fig. 4b) by the climatological easterly wind (Fig. 16a), yielding a reduction in negative MSE tendencies at the domain center in (Z2 in Fig. 15a). The horizontal structure of the Rossby waves yields zonal advection of MSE anomalies that generally leads the meridional advection of MSE anomalies. The lower- and midtropospheric climatological easterlies over and to the east of the domain center (Figs. 16a,b) also induce large-scale speed convergence, providing a favorable
environment for scale contraction and amplification of the Rossby waves via wave accumulation (e.g., Holland 1995; Sobel and Bretherton 1999; Sobel and Maloney 2000). Such a result suggests that the Rossby waves are initially triggered by the TC and subsequently amplified by wave accumulation induced by climatological large-scale lower- and midtropospheric easterly wind.

In aggregate, these results suggest that horizontal advection of MSE anomalies is driven by the interaction between the climatological mean environment and the TC and its associated Rossby waves. Specifically, zonal advection by the climatological lower- and midtropospheric easterly wind weakens the negative MSE anomalies at the domain center and strengthens negative MSE anomalies at the west region in the absence of Rossby waves, whereas both meridional and zonal advection by the Rossby waves also strengthen the negative MSE anomalies at the domain center and west region during Rossby wave passage. With an understanding of the horizontal advection of MSE anomalies, section 3f(2) focuses on the horizontal structure of the flux anomalies.

2) PHYSICAL INTERPRETATION OF FLUX ANOMALIES

Flux anomalies following TC passage are primarily dominated by two factors: 1) TC-induced negative SST anomalies and 2) Rossby waves. Negative flux anomalies occur immediately following TC passage, with the strongest flux anomalies concentrated near the domain center and extending ~2500 km to its northwest (F1 in Fig. 13). Further insight into the flux anomalies is obtained by partitioning the term into its two largest components: surface latent heat flux anomalies at days 10.5 (Fig. 17a) and 13.5 (Fig. 17b) and longwave heating rate anomalies at days 10.5 (Fig. 17c) and 13.5 (Fig. 17d). W2, LH1, and LW1 denote the TC-induced Rossby wave.
waves, salient regions of surface latent heat flux anomalies, and salient regions of longwave radiative heating rate anomalies, respectively, in Fig. 17. The flux anomalies at the domain center are primarily composed of surface latent heat flux anomalies (LH1 in Figs. 17a,b) with a relatively small contribution from longwave heating rate anomalies (LW1 in Figs. 17c,d), consistent with prior work (Sobel and Camargo 2005; Hart et al. 2007; Emanuel 2008; Schenkel and Hart 2011). In contrast, surface latent heat flux and longwave heating rate anomalies are comparable in the west region. Following day +10 after TC passage, there is a substantial decrease in both the magnitude and horizontal extent of the flux anomalies (F1 in Fig. 13f) as a result of decreases in surface latent heat flux anomalies (LH1 in Fig. 17b), consistent with Jullien et al. (2014). The reduction in negative surface latent heat flux anomalies is coincident with the reduction in boundary layer negative MSE and latent energy anomalies at the domain center (L in Figs. 9b,d, respectively). Additionally, the negative surface latent flux anomalies and longwave heating rate anomalies coupled with the negative horizontal advection of MSE anomalies suggest that all terms act to strengthen the negative MSE anomalies in the west region, perhaps explaining why these negative MSE anomalies are relatively larger in magnitude and longer lived compared to the domain center (Fig. 10).

To provide insight into the source of the surface latent heat flux anomalies, SST anomalies are shown for days +10.5 (Fig. 16e) and +13.5 (Fig. 16f). W2 and S1 denote the TC-induced Rossby wave and salient regions of SST anomalies, respectively, in Fig. 16. The correspondence between the negative surface latent heat flux anomalies (LH1 in Figs. 17a,b) and negative SST anomalies (S1 in Figs. 16e,f) suggests that the negative SST anomalies may force the surface latent heat flux anomalies, consistent with prior work (Jullien et al. 2014). However, the lack of strong agreement between the negative SST anomalies and surface latent heat flux anomalies (especially at day +13.5) suggests that equilibration occurs between the boundary layer moisture and reduced SSTs, decreasing the negative surface latent heat fluxes, similar to prior work (Jullien et al. 2014).

An analysis reveals that the longwave heating rate anomalies are primarily composed of TOA upwelling longwave radiative flux anomalies \( F_{\text{lw}}^{\text{toa}} \) with plan view plots of \( F_{\text{lw}}^{\text{toa}} \) shown for days +10.5 (Fig. 18a) and +13.5 (Fig. 18b). W2 and F1 denote TC-induced Rossby waves and salient regions of \( F_{\text{lw}}^{\text{toa}} \), respectively, in Fig. 18. \( F_{\text{lw}}^{\text{toa}} \) (LW1 in Fig. 18) is generally consistently and weakly positive compared to the other budget terms at the domain center and west region. However, \( F_{\text{lw}}^{\text{toa}} \) is enhanced during Rossby wave passage particularly on the western flank of the Rossby wave (W2 in Fig. 18). The positive \( F_{\text{lw}}^{\text{toa}} \) may be attributable to the reduced greenhouse gas effect of water vapor that results from tropospheric dry anomalies primarily induced by the TC as hypothesized by Emanuel (2008). These dry anomalies are strengthened by the negative MSE anomalies associated with the Rossby wave.

These results suggest that the negative surface latent heat flux anomalies (especially at the domain center), longwave heating rate anomalies, and horizontal advection of MSE anomalies each play an important role in generating negative MSE anomalies. These results also emphasize the importance of the broad spatiotemporal feedbacks of the TC-induced negative SST anomalies upon the full troposphere. However, questions remain
regarding the representativeness of the budgets on the days selected, which motivates a discussion of the temporal evolution of the MSE budget time tendencies at the domain center and west region in section 3g.

g. Temporal structure of MSE budget time tendency anomalies

To depict the time scales of MSE budget time tendencies, time series of area-averaged MSE budget time tendency anomalies at the domain center (Fig. 19a) and west region (Fig. 19b) are presented for horizontal advection of MSE anomalies and flux anomalies, respectively, and are referenced in the text. The black vertical lines denote the passage of the TC and the TC-induced Rossby waves (labeled with W1–W6) and are referenced in the text.

FIG. 19. Time series of area-averaged vertically integrated MSE budget anomalies (W m$^{-2}$) at the (a) domain center and (b) west region during typhoon passage. The labels H1–H2 and F1–F2 denote salient periods of horizontal advection of MSE anomalies and flux anomalies, respectively, and are referenced in the text. The black vertical lines denote the passage of the TC and the TC-induced Rossby waves (labeled with W1–W6) and are referenced in the text.

H1–H2 and F1–F2 denote salient periods of horizontal advection of MSE anomalies and are referenced in the text. At the domain center, the negative flux anomalies are dominant in the first two weeks following TC passage (F1 in Fig. 19a) before steadily decreasing to relatively smaller values (F2 in Fig. 19a). In contrast, horizontal advection of MSE anomalies at the domain center begins as a positive tendency in the first month following TC passage (H1 in Fig. 19a) before eventually becoming negative (H2 in Fig. 19a). In the west region, the flux anomalies (F2 in Fig. 19b) and horizontal advection of MSE anomalies are comparable in magnitude for the first month following TC passage, after which the latter term dominates.

A time series of the zonal and meridional components of horizontal advection of MSE anomalies for the domain center (Fig. 20a) and west region (Fig. 20b) is shown next. H1–H2 denote salient periods of horizontal

FIG. 20. As in Fig. 19, but for partitioned zonal and meridional advection of MSE anomalies (W m$^{-2}$). The labels H1–H2 denote salient periods of horizontal advection of MSE anomalies and are referenced in the text.
advection of MSE anomalies at the domain center in Fig. 20. Figure 20a reveals that the zonal advection of MSE anomalies at the domain center is generally characterized by positive MSE tendencies, whereas the west region (Fig. 20b) is generally characterized by periodic negative MSE tendencies due to Rossby waves. However, the negative meridional advection of MSE anomalies associated with the Rossby waves (Fig. 20a) eventually becomes dominant at the domain center because of the weakening of the zonal MSE gradient at the domain center (transition from H1 to H2 in Fig. 20a). In the west region, horizontal advection of MSE anomalies is generally characterized by negative meridional and, to a lesser extent, zonal advection of MSE anomalies associated with Rossby waves throughout the two months following TC passage.

The time series of the area-averaged surface latent heat flux and longwave radiative heating components of the flux anomalies are featured at the domain center (Fig. 21a) and west region (Fig. 21b). The F1, LH1–LH2, and LW1 denote salient periods of flux anomalies, surface latent heat flux anomalies, and longwave heating rate anomalies, respectively, in Fig. 21. Consistent with prior work (Jullien et al. 2014), Fig. 21a reveals that surface latent heat flux anomalies at the domain center peak shortly following TC passage (LH1 in Fig. 21a) before steadily decreasing in the following several weeks (LH2 in Fig. 21a). In contrast, surface latent heat flux anomalies at the west region are generally steady over the same period (F1 in Fig. 21b). The surface latent heat flux anomalies at the domain center and west region do not exhibit substantial sensitivity to the Rossby waves. It is important to note that the surface latent heat flux anomalies eventually return to climatology ~40 days after TC passage in the domain center and west region, consistent with prior work (Jullien et al. 2014). Lastly, comparison of the time–longitude plots of SST anomalies (Fig. 2) with the time series of surface latent heat flux anomalies (Fig. 21) reveals a restoration of the former well before the latter, consistent with prior work (Jullien et al. 2014) and suggesting that an equilibration occurs between boundary layer moisture and reduced SSTs at the domain center and west region.

In contrast to the surface latent heat flux anomalies, the longwave heating rate anomalies are generally consistently negative at both the domain center and west region for ~40 days following TC passage. It is important to note that the restoration of the longwave heating rate and surface latent heat flux anomalies back to climatology at both the domain center and west region occurs nearly simultaneously, perhaps suggesting that the latent heat flux anomalies force the longwave heating rate anomalies (Fig. 21). As previously mentioned, the longwave heating rate anomalies at the domain center and west region exhibit sensitivity to the Rossby waves (Fig. 21). Finally, it should be emphasized that the long-lived negative MSE anomalies at the west region occur owing to the absence of positive MSE budget time tendency terms.

Consistent with the prior section, these results confirm that the MSE budget time tendencies at the domain center and west region are driven by a combination of the TC-induced negative SST anomalies and TC-induced Rossby waves. In light of this, Fig. 21 conservatively suggests that TC environmental impacts, as measured by the time scales of the TC-induced negative SST anomalies and Rossby waves, occur for ~40 days following TC passage. Beyond this time, it is difficult to discriminate whether the negative MSE anomalies are forced by the TC.

4. Summary and discussion

The present study has quantified the spatiotemporal scales of environmental tropospheric thermodynamic
anomalies before and after WPAC TC passage and attributed these anomalies to processes using vertically integrated MSE budgets. The salient results of the present study are summarized in a schematic in Fig. 22. The primary conclusion of the present study is that large-scale tropospheric negative MSE anomalies generally only occur for larger TCs for two months following TC passage owing to a combination of the TC, TC-induced Rossby waves, and the MJO. Specifically, storm-relative composites of reanalyses reveal that vertically integrated negative MSE anomalies occur for two months following TC passage in a region extending from the TC track to 3000 km west of the TC track. The negative MSE anomalies are modulated by a series of westward-propagating negative MSE anomalies (particularly in the typhoon composite) occurring throughout the 60 days following TC passage. An alternate set of composites for typhoons stratified according to whether the large-scale environment is favorable or unfavorable for TC-induced Rossby wave dispersion reveals that the westward-propagating MSE anomalies are likely due to TC-induced Rossby waves radiated to the south and east of the TC. More importantly, the composite of typhoons with favorable environments for TC-induced Rossby wave dispersion exhibits substantially stronger negative MSE anomalies following typhoon passage, while the composite of typhoons with unfavorable environments for Rossby wave dispersion does not exhibit any significant negative MSE anomalies. The substantial difference in MSE anomalies between favorable and unfavorable TC-induced Rossby wave dispersion composites is likely due to larger TCs in the favorable composite triggering broader, stronger negative SST anomalies and exciting TC-induced Rossby waves as summarized in Fig. 22. Such a result suggests that only a subset of larger TCs is responsible for forcing negative MSE anomalies following TC passage (Fig. 22).

Consistent with prior work, the negative vertically integrated MSE anomalies are primarily composed of lower- and midtropospheric negative latent energy anomalies with a smaller contribution from surface and upper-tropospheric negative sensible heat anomalies. The negative MSE anomalies can be further split into two regions based upon their proximity to the TC track: 1) the domain center and 2) the west region (1100 km west of the domain center). Both regions contain boundary layer negative latent energy and sensible heat anomalies, which suggests that these anomalies are forced by an anomalous reduction in sea surface fluxes induced by the TC. Midtropospheric negative latent energy anomalies are strongest during Rossby wave passage at both the domain center and west region, suggesting that these anomalies are forced by the Rossby waves. The upper-tropospheric negative sensible heat anomalies in both regions are likely partially due to a convectively suppressed MJO, although a contribution from the TC cannot be ruled out. It is important to note that negative MSE anomalies are consistently stronger at the west region compared to the domain center.
Vertically integrated MSE budgets confirm that the negative MSE anomalies at the domain center are primarily forced by a reduction in surface latent heat flux anomalies induced by the negative SST anomalies forced by the TC. Negative MSE tendencies at the domain center are also induced by enhanced $F_{lw}^{toa}$ potentially caused by a reduction in the greenhouse gas effect of water vapor, while negative zonal and meridional advection of MSE anomalies induced by the Rossby waves also yield negative MSE tendencies. The budget analysis in the west region is generally similar except that all terms are comparable in magnitude and relatively weaker (except $F_{lw}^{toa}$). The time scales of the Rossby waves and surface latent heat flux anomalies induced by the TC suggest that TCs can impact their synoptic-scale tropospheric environment for ~40 days following TC passage. Beyond this time, the link between the TC and negative MSE anomalies is more tenuous given that TCs do not occur in isolation from other phenomena.

Together with prior work (Sobel and Camargo 2005; Hart et al. 2007, 2008; Schenkel and Hart 2011), the results of the present study suggest that only larger TCs can stabilize their thermodynamic environment on spatiotemporal scales extending well beyond the TC. While the present study has addressed many questions concerning the impacts of TCs upon their atmospheric environment, there remain additional questions that have been raised as a result of the present study:

- How does the atmospheric environment respond to the impacts of TCs in the absence of the Rossby waves and the MJO?
- Does the aggregate impact of the cooling and drying of the troposphere by WPAC TCs help constrain the seasonal cycle of MSE within the WPAC tropics?
- Can the results of the present study for WPAC TCs be extended to other basins (e.g., North Atlantic and eastern North Pacific) given the differences in the large-scale environment and in the time scales of the environmental atmospheric response in other basins (Hart et al. 2007)?

These questions remain the focus of ongoing research.

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