On the Convective Coupling and Moisture Organization of East Pacific Easterly Waves

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ABSTRACT

Processes associated with the local amplification of easterly waves (EWs) in the east Pacific warm pool are explored. Developing EWs favor convection in the southwest and northeast quadrants of the disturbance. In nascent EWs, convection favors the southwest quadrant. As the EW life cycle progresses, convection in the northeast quadrant becomes increasingly prominent and southwest quadrant convection wanes. The EW moisture budget reveals that anomalous meridional winds acting on the mean meridional moisture gradient of the ITCZ produce moisture anomalies supportive of convection in the southwest quadrant early in the EW life cycle. As EWs mature, moisture anomalies on the poleward side of the EW begin to grow and are supported by the advection of anomalous moisture by the mean zonal wind.

In the southwest and northeast portions of the wave, where convection anomalies are favored, lower-tropospheric vorticity is generated locally through vertical stretching that supports a horizontal tilt of the wave from the southwest to the northeast. EWs with such tilts are then able to draw energy via barotropic conversion from the background cyclonic zonal wind shear present in the east Pacific. Convection anomalies associated with EWs vary strongly with changes in the background intraseasonal state. EWs during westerly and neutral intraseasonal periods are associated with robust convection anomalies. Easterly intraseasonal periods are, at times, associated with very weak EW convection anomalies because of weaker moisture and diluted CAPE variations.

1. Background

East Pacific easterly waves (EWs) are off-equatorial convectively coupled phenomena that occur during boreal summer. EWs provide a majority of the tropical convective variance at time scales less than 10 days (Kiladis et al. 2006), are associated with 25%–40% of deep convective clouds in the ITCZ (Tai and Ogura 1987; Gu and Zhang 2002), and seed the majority of east Pacific tropical cyclones (e.g., Avila and Guiney 2000; Avila et al. 2003; Pasch et al. 2009). East Pacific EWs are energized from the latent heat release associated with convection and by extracting kinetic energy from the background flow (Tai and Ogura 1987; Maloney and Hartmann 2001; Hartmann and Maloney 2001; Serra et al. 2008; Aiyyer and Molinari 2008; Serra et al. 2010; Rydbeck and Maloney 2014; Hoover 2015). In this study, we show that convection within the EW is also able to modulate the horizontal orientation of an EW’s circulation in addition to its intensity.

The orientation, or tilt, of the EW is critical to the barotropic conversion of mean kinetic energy to perturbation kinetic energy. In the presence of cyclonic shear of the mean zonal wind, as typically occurs in the east Pacific warm pool, EWs are energized if the wave’s horizontal flow field is elongated along a diagonal from southwest (SW) to northeast (NE). Such EWs produce a downdraft momentum flux in the presence of cyclonic shear of the mean zonal wind associated with eddy growth at the expense of the mean flow. The barotropic conversion term in the perturbation kinetic energy budget of EWs responsible for this process is $\frac{\partial}{\partial t} \frac{\mathbf{u} \cdot \mathbf{V}}{\partial y}$ (Rydbeck and Maloney 2014). In this study, we explore the role of convection in producing the observed SW–NE horizontal tilt of EWs in the east Pacific.

Since the discovery of EWs, different conceptual models of the wave’s convection and circulation have been hypothesized (e.g., Dunn 1940; Riehl 1945, 1948, 1954; Palmer 1952; Malkus and Riehl 1964). In particular, many analyses have suggested various phasing of convection with regard to the wave axis. We briefly...
summarize three leading conceptual models of EW phasings: the Caribbean, the Atlantic, and the African wave models.

The Caribbean wave model, often referred to as Riehl’s classical EW model, was derived from four-times-daily atmospheric soundings over Puerto Rico and visual observations (Riehl 1945, 1948, 1954). Riehl’s work built on findings by Dunn (1940) regarding the relationship of precipitation and cloud types to “isallobaric centers” (regions of greatest negative pressure tendency) in the Caribbean. In the Caribbean wave model, to the west of the wave axis, the depth of the moist layer is at a minimum, near 5000 feet (ft; ~1500 m), and is largely associated with nonprecipitating trade cumulus and cumulus humilis clouds. This region of subsidence and surface divergence precedes the wave axis. Near and just to the east of the wave axis, the moist layer deepens rapidly to above 20000 ft (~6100 m), and cumulus congestus, cumulonimbus, altostratus, stratocumulus, altocumulus, and cirrus clouds are prevalent. Far to the east of the wave axis, the depth of the moist layer gradually decreases but is still accompanied by cumulus, cumulus congestus, stratocumulus, altostratus, cirrus, and occasional cumulonimbus clouds. Thus, in the Caribbean wave model, deep convection and column-integrated moisture anomalies favor regions in and behind the wave axis. These observations of surface convergence to the east and surface divergence to the west of the wave axis were hypothesized by Riehl (1954) to result from the principles of conservation of potential vorticity in the moist layer for easterly currents passing through the wave. Subsequent modeling studies of EWs in the Caribbean corroborated Riehl’s observational findings (Krishnamurti and Baumhefner 1966; Yanai and Nitta 1967).

The Atlantic wave model, often referred to as Frank’s inverted V model, was first proposed by Simpson et al. (1968) and later developed by Frank (1969) using cloud mosaics from the Television and Infrared Observation Satellite (TIROS) over the Atlantic. The distribution of clouds in the Atlantic wave model is banded and forms an upside-down V shape. The Atlantic wave model mainly differs from the Caribbean wave model in that convection is equally distributed across the wave axis. The deep convective features in the Atlantic wave model are present both upstream and downstream of the wave axis and do not display a rapid buildup in cloud depth near the axis as in the Caribbean wave model. Frank (1969) suggested that differences in the mean state and the vigor of waves sampled in the studies were at least partly responsible for the differing distributions of convection between the Atlantic and Caribbean wave models.

The African wave model proposed by Carlson (1969) suggested that the maximum convection and precipitation associated with EWs over West Africa occurred ahead of the wave axis. Using GATE observations and Synchronous Meteorological Satellite (SMS) IR brightness temperatures, Reed et al. (1977) was able to confirm Carlson’s (1969) results. The peak upward vertical motion occurred in advance of the wave axis and maximized near 700 hPa. The upstream convection persisted even when opposed by a relatively dry lower troposphere transported southward from the Sahara, suggesting that large-scale forcing was aiding convection ahead of the wave. The phasings of convection relative to the wave axis vary considerably based on the season and geographic location, but Carlson’s original findings generally hold for waves in West Africa and south of the African easterly jet axis during June–October (Gu et al. 2004; Kiladis et al. 2006 and references therein).

Recent studies of EWs in the east Pacific suggest various phasings of convection and moisture relative to the wave axis. Analysis of EW moisture by Serra et al. (2008) in the east Pacific agrees with the Caribbean wave model. Positive specific humidity anomalies are confined to the lower troposphere in advance of the wave axis and rapidly deepen near the wave axis. These results also agree with an earlier study by Tai and Ogura (1987) that suggested deep convection occurs within or slightly behind the wave axis in the east Pacific. Yet results from Petersen et al. (2003) using Doppler radar from EPIC-2001 suggest that EWs in the east Pacific are likely to have deep convection in front of the wave axis and weaker but more widespread convection at and behind the wave axis, although this study was derived for a relatively short time record. Furthermore, Serra and Houze (2002) note that EWs in the east Pacific are readily identified in the east Pacific by their inverted V shapes noted on visible satellite imagery, as in the Atlantic wave model. These differences in the phasing of convection with the wave circulation may partly result from sampling the wave at various stages of its life cycle and/or during different background states of the east Pacific. The analysis below will support this contention.

Tropical convection shows strong sensitivity of precipitation to the buildup and removal of free-tropospheric moisture (Raymond 2000; Sobel et al. 2004; Bretherton et al. 2004; Peters and Neelin 2006; Holloway and Neelin 2009; Sahany et al. 2012). We will show that anomalous convection favors certain regions of EWs, and these regions are also areas where vertically integrated moisture anomalies occur. The regulatory processes for the moisture anomalies are examined using a column-integrated moisture budget. We also partition the budgets with respect to intraseasonal phase, which is known to strongly modulate the background fields through which EWs propagate and are energized (e.g., Nitta and Takayabu 1985;
Molinari and Vollaro 2000; Maloney and Hartmann 2001; Aiyyer and Molinari 2008; Rydbeck and Maloney 2014). Understanding the preferred regions of anomalous convection within EWs is important, as the convection can locally generate vorticity through vorticity convergence. We hypothesize that the local generation of vorticity by convection leads to a horizontally tilted structure of the EW favorable for wave intensification in regions of background cyclonic shear.

A description of the data used in the study and the methods for compositing both EWs and intraseasonal events in the east Pacific are described in section 2. The evolution of strong EWs during various intraseasonal states is described in section 3. The column-integrated moisture budget for EWs is examined in section 4. The role of adiabatic forcing to support EW convection is investigated in section 5. Section 6 discusses how EW convection modulates the horizontal wave structure by briefly examining the vorticity budget. Section 7 presents the conclusions of the study.

2. Data and methods

Temperature, specific humidity, precipitation, surface evaporation, and winds from the European Centre for Medium-Range Weather Forecasts interim reanalysis (ERA-Interim; Dee et al. 2011) are used throughout the study. The 6-hourly, 1.5° grid-spaced data cover a 20-yr period from 1991 to 2010. For the purposes of calculating column-integrated moisture budgets, we use raw variables that span 1000–200 hPa with a 50-hPa interval. The top level (200 hPa) is quite close to the typical level of neutral buoyancy for a boundary layer parcel during boreal summer in the Intra-American Seas (Jordan 1958).

As a proxy for convection, the newly released National Climatic Data Center (NCDC) outgoing longwave radiation (OLR) daily climate data record from 1991 to 2010 is used (Lee 2014; NOAA 2014). Because only daily averages are available, the 1° × 1° grid-spaced OLR data are linearly interpolated to 6-hourly time steps to be utilized in the EW composites shown later. The linearly interpolated NCDC OLR dataset is largely consistent with 6-hourly linearly interpolated to 6-hourly time steps to be utilized in the EW composites (Rydbeck and Maloney 2014).

EW fields are compositcd both as a function of EW phase and intraseasonal phase. The framework for identifying strong EWs relies on the use of the leading empirical orthogonal function (EOF) pair of 2.5–12-day bandpass-filtered 700-hPa vorticity anomalies. Tropical cyclone–related vorticity anomalies are removed as in Aiyyer et al. (2012) before the EOF analysis is performed on the filtered vorticity anomalies. The leading EOF pair describes a northwestward-propagating vorticity anomaly that is used to partition the EW’s life cycle into eight EW phases. Only strong EWs, those exceeding 0.5 standard deviations, are chosen using amplitudes derived from the corresponding principal component (PC) pair. More details regarding the methods used to calculate both EW and intraseasonal phase angle, amplitude, and selection criteria, as well as the sensitivities of the methodology, are described in Rydbeck and Maloney (2014).

The phase and amplitude criteria of the intraseasonal oscillation in the east Pacific are based on the normalized leading PC of 30–90-day bandpass-filtered 850-hPa zonal wind anomalies. Local maxima greater than 1.0 standard deviation are defined as the peak of westerly intraseasonal events. Local minima less than −1.0 standard deviation are defined as the peak of easterly intraseasonal events. Neutral intraseasonal events are defined as the time step closest to zero for values between −0.5 and 0.5 standard deviations. Each intraseasonal period is defined to include the 5 days before and after the event’s peak. As in Rydbeck and Maloney (2014), the results are not overly sensitive to reasonable changes in the thresholds used to identify EWs and intraseasonal events.

3. Easterly wave evolution

The background state through which east Pacific EWs propagate shows dramatic variations in winds and shear, moisture and convection, and SST and latent heat fluxes on intraseasonal time scales (Maloney and Hartmann 2000, 2001; Molinari and Vollaro 2000; Maloney and Kiehl 2002; Maloney and Esbensen 2003, 2005, 2007; Aiyyer and Molinari 2008; Maloney et al. 2008; Rydbeck et al. 2013; Crosbie and Serra 2014). Westerly intraseasonal periods are associated with enhanced low-level westerlies, precipitation, and surface latent heat fluxes. Easterly intraseasonal periods are associated with anomalous low-level easterlies, suppressed precipitation, and suppressed surface latent heat fluxes. These variations are associated with differences in the generation of EW kinetic energy, with westerly intraseasonal periods characterized by significant increases in EW kinetic energy versus easterly intraseasonal periods (Maloney and Hartmann 2001; Hartmann and Maloney 2001; Aiyyer and Molinari 2008; Rydbeck and Maloney 2014; Crosbie and Serra 2014). To understand the intraseasonal variability of EWs as well as the evolution of EW convection and circulation anomalies, we review their composite structure as a function of intraseasonal phase.

Figure 1 shows the composite evolution of EW 2.5–12-day bandpass-filtered 700-hPa vorticity and OLR anomalies during neutral intraseasonal periods. The EW initially develops in association with a weak positive vorticity
anomaly located west of Panama in phases 7, 8, and 1. Throughout this description of EW evolution, we concentrate on the half wavelength encompassing the positive vorticity anomaly and use this synonymously with the term “easterly wave.” This positive anomaly intensifies as OLR anomalies become increasingly negative. By phase 2, negative OLR anomalies associated with the incipient vortex are most prominent in the SW quadrant of the growing wave. In phase 4, the nascent positive vortex is tilted from SW to NE, and negative OLR anomalies are still largely confined to the SW quadrant of the wave. Such a horizontal tilt could arise from low-level vorticity convergence forced by deep convection in that same SW quadrant. We will explore this stretching mechanism below when analyzing the vorticity budget. By phase 5, the vortex is prominently tilted from SW to NE, which would support EW kinetic energy generation by barotropic conversion in a region of cyclonic shear of the zonal wind.

**FIG. 1.** The 2.5–12-day bandpass-filtered 700-hPa relative vorticity ($\times 10^{-6}$ s$^{-1}$; line contours) and OLR (W m$^{-2}$; color contours) anomalies composited for neutral intraseasonal periods and strong easterly waves. The vorticity contour interval is $1 \times 10^{-6}$ s$^{-1}$. Zero-relative vorticity contours are thick solid lines, and positive (negative) contours are solid (dashed). Phase and number of days included in each composite are shown on the right above each panel.
that characterizes the basic-state flow in this region [e.g., Fig. 5 in Rydbeck and Maloney (2014)]. During phase 5, OLR minima are present in both the SW and NE quadrants of the wave. The wave continues to intensify in phases 6, 7, and 8 as the OLR minimum in the SW quadrant weakens and the minimum previously in the NE quadrant intensifies and shifts slightly west of the wave. The intensification of the vorticity maximum and OLR minimum persist until the EW encounters cooler waters to the northwest of the east Pacific warm pool.

EWs during westerly intraseasonal periods are characterized by stronger vorticity and OLR anomalies (Fig. 2) than waves during neutral intraseasonal periods. Composites of EW anomalies during westerly intraseasonal periods are noisier than those during neutral intraseasonal periods, likely because of the reduced number of events. The nascent vortex to the west of the Panama Bight in phase 8 quickly intensifies by phase 1, while the anomalous OLR minimum in the SW quadrant increases in magnitude from $-6$ to $-10$ W m$^{-2}$.

During phases 1 and 2, the OLR minimum in the SW quadrant of the nascent wave is connected to a tongue of negative OLR anomalies extending westward from Panama. Like EWs during neutral intraseasonal periods, the

**Fig. 2.** As in Fig. 1, but for strong easterly waves during westerly intraseasonal periods.
OLR minimum is largely confined to the SW quadrant as the wave matures. As the wave continues to grow in phase 5, a second OLR minimum develops in the NE quadrant. The wave axis also exhibits a prominent SW–NE tilt at this time. Through phases 6, 7, and 8, the OLR minimum in the NE quadrant intensifies from $\sim 8$ to less than $\sim 16 \text{ W m}^{-2}$. The OLR minimum during these phases slightly leads the EW. The vortex eventually weakens in association with a weakened OLR minimum upon entering a region of climatologically cooler sea surface temperatures.

Composite EW vorticity is weaker during the incipient phases for easterly intraseasonal periods and slower to develop compared to neutral and westerly intraseasonal periods (Fig. 3). OLR minima associated with the EW do not display a consistent strengthening in the early stages of the EW life cycle (phases 1, 2, 3, and 4). Unlike neutral and westerly intraseasonal periods, the OLR minimum slightly lags the positive vortex such that convection anomalies are located in the SE quadrant of the nascent vortex during phases 1 and 2. OLR anomalies associated with the nascent vortex weaken by phases 3 and 4 to greater than $\sim 4 \text{ W m}^{-2}$. OLR minima favor the NE quadrant by phases 6, 7, and 8. These phases are also when the maturing EW vortex exhibits a

*Fig. 3.* As in Figs. 1 and 2, but for strong easterly waves during easterly intraseasonal periods.
more defined SW–NE tilt. Vorticity and OLR anomalies associated with the mature EW are largely collocated during phases 1, 2, and 3.

The relationship of convection to the circulation of EWs varies as the wave grows from an incipient disturbance to a robust wave. The convection displays noteworthy zonal and meridional asymmetries. Generally, the OLR minimum prefers the SW quadrant of the wave early in the life cycle and favors the NE quadrant as the wave reaches maturity. Strong convective anomalies are simultaneously present in both the SW and NE quadrants of growing waves during neutral and westerly intraseasonal periods. The asymmetries in convection occur in conjunction with asymmetries in the circulation: namely, a SW–NE tilt of the horizontal flow field. We later discuss how the convection and circulation may be coupled to produce such an effect. The exception to this generalization occurs during easterly intraseasonal periods when convection is much weaker during the formative stages. Eventually a single but broad OLR minimum is centered near the vortex during peak intensity. The exact phasing and vigor of convection relative to the EW circulation varies as a function of the intraseasonal background state, with the most vigorous EWs occurring during westerly intraseasonal periods and weakest waves during easterly intraseasonal periods. To verify that the features are strongly related to EWs and not tropical cyclones that project onto the same frequency band, we also removed the influence of tropical cyclones from the anomaly fields in the same manner as Rydbeck and Maloney (2014). The same conclusions hold (not shown here).

4. Moisture budget

As mentioned previously, tropical convection exhibits strong sensitivity to free-tropospheric moisture. More specifically, Raymond et al. (1998) noted the importance of moistening the middle troposphere to the development and maintenance of deep convection in east Pacific EWs during the summer of 1991. EW composites of 2.5–12-day bandpass-filtered total precipitable water (TPW), OLR, and 700-hPa wind anomalies during phases 3 and 7 for neutral, westerly, and easterly intraseasonal periods, respectively, are shown in Fig. 4 to determine if a similar relationship between moisture and convection is present. Regardless of the intraseasonal period, TPW maxima (minima) are generally collocated and supportive of the OLR minima (maxima) during both incipient and mature stages of the EW. Building on the previous large body of observational and modeling evidence that suggests daily mean tropical convection and precipitation are strongly constrained by column moisture, we now analyze the processes that regulate moisture locally on EW time scales utilizing the vertically integrated moisture budget. It is hypothesized that moisture anomalies in the SW and NE quadrants support convection anomalies that help tilt growing waves in a manner favorable for increased barotropic conversion of EW kinetic energy, as quantified in Rydbeck and Maloney (2014). The vertically integrated moisture budget is approximately given by the following:

$$\left[ \frac{\partial q}{\partial t} \right] = - [u_h \cdot \nabla_h q] - \left[ \frac{\partial q}{\partial \rho} \right] + E' - P',$$  

where the brackets represent the mass-weighted vertical integral from 1000 to 200 hPa, and the prime represents a 2.5–12-day bandpass filter. The term on the left-hand side of Eq. (1) represents the vertically integrated 2.5–12-day bandpass-filtered specific humidity tendency. This term is simply referred to as the EW moisture tendency. The first term on the right-hand side of Eq. (1) is the moisture tendency resulting from horizontal advection. The second term on the right-hand side is the moistening by vertical advection. The third and fourth terms on the right-hand side are the column moisture tendency as a result of surface evaporation and precipitation, respectively.

a. Neutral intraseasonal periods

The composite evolution of the EW moisture tendency for phases 3 and 7 during neutral periods is shown in the top row of Fig. 5. The positive moisture tendency during the early period of the EW life cycle maximizes in the SW quadrant of the wave, near the transition between the ridge and trough. That same positive moisture tendency is present in the SE quadrant of the ridge before the nascent EW is well established, as can be seen in phase 7 near 5°N, 95°W. For the strongest EW vorticity center seen in phase 7, the positive moisture tendency is elongated and leads the wave with maxima in advance of the SW and NE quadrants. Compared with Fig. 4, anomalous moistening is occurring in advance of the negative OLR and positive TPW anomalies associated with the westward-propagating wave. By phases 8, 1, and 2 (not shown), the positive moisture tendency in advance of the NE quadrant of the mature EW intensifies, and the local maximum in the SW quadrant relaxes such that only a single moistening maximum occurs in advance of the wave, as can be seen in phase 3. Peak moistening over the life cycle of the wave first occurs ahead of the SW quadrant, with values of 3.5 mm day⁻¹ during phase 3 and in advance of the NE quadrant of developed waves at approximately 4 mm day⁻¹ during phases 1 and 2 (not shown). In the interim stages, maxima in moistening occur simultaneously in advance of both the SW and NE.
quadrants during phase 7, with values of 2.5 mm day\(^{-1}\). For growing waves in phases 3 and 7, moisture tendency minima are located behind the wave axis, in the SE quadrant, indicating drying behind the wave.

Early in the wave’s life cycle during neutral periods, evaporation opposes the positive moisture tendency by up to \(-0.7\) mm day\(^{-1}\) in the SW quadrant (not shown). A similar evaporation tendency is observed for EWs during both westerly and easterly intraseasonal periods. The combination of vertical moisture advection and precipitation is described next. In general, separating cause and effect between condensational heating and vertical velocity in the tropics is complicated (e.g., Sobel and Bretherton 2000; Back and Bretherton 2009); hence, we choose to examine these two processes in tandem. Confidently calculating vertical advection minus precipitation using reanalysis data is difficult, and, as a result, the difference is formulated as the residual of the other terms in the budget. Any analysis increments used to constrain the ERA-Interim moisture field in the assimilation process are also likely the result of errors in moist physics that would be included in the sum of vertical advection and precipitation (e.g., Wolding and Maloney 2015); hence, it makes sense to calculate this difference as a residual. We note that vertical advection minus precipitation calculated explicitly is qualitatively similar to vertical advection minus precipitation calculated as the residual of the other terms, however. In the difference between the vertical advection and precipitation (bottom row of Fig. 5), positive regions indicate where vertical advection is moistening the atmosphere in opposition to drying by precipitation. Thus, moistening during the incipient stages of the EW not counteracted by precipitation preferentially occurs.

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**FIG. 4.** The 2.5–12-day bandpass-filtered TPW (mm; color contours), OLR (W m\(^{-2}\); line contours), and 700-hPa wind (m s\(^{-1}\); vectors) anomalies composited for phases (left) 3 and (right) 7 of (top) neutral, (middle) westerly, and (bottom) easterly intraseasonal periods and strong easterly waves. The OLR contour interval is 3 W m\(^{-2}\). Zero-OLR contours are thick solid lines, and negative (positive) contours are solid (dashed). Phase and number of days included in each composite are shown on the right above each panel. Reference wind vectors are located in the upper right of each panel. Wind vectors less than 0.3 m s\(^{-1}\) are not shown.
ahead of the wave, with values of 0.5–1 mm day\(^{-1}\). Anomalous drying resulting from precipitation that is stronger than vertical advection occurs behind the wave. Processes that might play a moistening role are adiabatic forcing and/or shallow nonprecipitating convection. We explore the former possibility in section 6 by estimating \(Q\)-vector divergence. However, the difference between vertical advection and precipitation is substantially smaller than the total moisture tendency.

Horizontal advection is the largest contributor to the positive tendency of column-integrated moisture tendencies ahead of EW convection (middle row of Fig. 5). The minima and maxima of column-integrated horizontal advection are collocated with and of comparable values to the column-integrated moisture tendency. Since precipitation and vertical advection largely cancel and evaporation weakly opposes the positive moisture tendencies of incipient waves, the overwhelming majority of the moisture tendency is due to horizontal advection. The contributions by zonal and meridional advection comparably contribute to the horizontal advection during neutral intraseasonal periods (not shown). The horizontal advection term can further be understood using the following decomposition of the horizontal advection to highlight the respective roles of the mean and perturbation flows and moisture gradients:

\[
-\nu_h \cdot \nabla_h q = \left[ \frac{\partial q}{\partial x} \right]' \nu' - \left[ \frac{\partial q}{\partial y} \right]' \nu' - \left[ \frac{\partial \bar{q}}{\partial x} \right]' \bar{\nu} - \left[ \frac{\partial \bar{q}}{\partial y} \right]' \bar{\nu} - \left[ \frac{\partial q''}{\partial x} \right]' \bar{\nu}' - \left[ \frac{\partial q''}{\partial y} \right]' \bar{\nu}' ,
\]  

(2)

where the overbar indicates a 10-day running mean, and the double prime indicates deviations from a 10-day
running mean. The first and second terms on the right-hand side (RHS) of Eq. (2) are the zonal and meridional advection of mean moisture by the perturbation zonal and meridional winds, respectively. The third and fourth terms on the RHS are the zonal and meridional advection of the perturbation moisture by the mean zonal and meridional winds, respectively. The fifth and sixth terms on the RHS are the advection of perturbation moisture by perturbation winds.

Of the six terms on the RHS of Eq. (2), the advection of mean moisture by the perturbation meridional winds \(-[\bar{u} \partial \bar{q} / \partial y]\); bottom row of Fig. 6) and the advection of perturbation moisture by the mean zonal winds \(-[\bar{u} \partial \bar{q} / \partial x]\); top row of Fig. 6) are the major contributors to the observed horizontal moisture advection seen in Fig. 5. The moistening resulting from advection of mean moisture by the perturbation meridional winds occurs preferentially ahead of the wave in the region of the wave containing a strong northerly wind component (see Fig. 4). This relationship reverses when the meridional moisture gradient flips sign as the wave is exiting the NW side of the warm pool near Baja California Sur. Figure 7 (bottom panel) shows the mean distribution of TPW during neutral intraseasonal periods. A zonally elongated strip of maximum precipitable water extends from the Panama Bight to the west-northwest to approximately 100°W, where it continues to the west along 9°N. As the incipient wave grows, it preferentially advects high TPW southward toward regions of lower TPW, creating anomalously high TPW values in the SW quadrant of the wave. The meridional advection of mean moisture by perturbation meridional winds contributes over 3 mm day\(^{-1}\) to the moisture tendency during the growing stages of the EW, as seen in phase 3 of Fig. 6.

Zonal advection of perturbation moisture anomalies by the mean zonal winds contributes over 1.5 mm day\(^{-1}\) to the moisture tendency increase in the SW quadrant of nascent EWs. In mature EW stages, zonal advection of perturbation moisture by the mean zonal winds moistens in excess of 2 mm day\(^{-1}\) ahead of convection in the NE quadrant. Perturbation zonal winds acting on the mean zonal moisture gradient \(-[\bar{u} \partial \bar{q} / \partial x]\) also contribute to moistening in this region by 0.5 mm day\(^{-1}\) (not shown).

To summarize, TPW anomalies are in phase with OLR anomalies in EWs. Regions ahead of the EW convection are moistened largely as a result of horizontal advection. During the early stages of an EW, moisture increases occur in the SW quadrant as a result of EW meridional winds acting on the mean moisture gradient collocated with the mean ITCZ. As the wave intensifies and shifts northwestward away from the strong meridional moisture
gradient on the equatorward side of the ITCZ, the positive moisture tendency in the SW quadrant weakens. At this time, moistening in advance of convection in the NE quadrant of the wave intensifies because of the zonal advection of moisture anomalies.

b. Westerly and easterly intraseasonal periods

The locations of column-integrated moisture tendency maxima for EWs during westerly and easterly intraseasonal periods are similar to those in the neutral intraseasonal composite relative to the EW train (top rows of Figs. 8 and 9). However, differences exist in their magnitude, especially during westerly intraseasonal periods. For example, the composite moisture tendency ahead of the mature EW in phase 7 during westerly intraseasonal periods is 40% stronger than the maximum during neutral intraseasonal periods. Likewise, the tendency of moisture in the SW quadrant of the nascent EW in phase 8 (not shown) is 40% stronger during westerly versus neutral intraseasonal periods. Figures 1 and 2 showed that the OLR anomalies during westerly intraseasonal periods increase more rapidly than those during neutral intraseasonal periods. Moisture tendencies acting to increase moisture anomalies that are supportive of the convection anomalies exhibit the same behavior.

EWs during easterly intraseasonal periods are characterized by weak and sometimes absent OLR anomalies during early stages of the EW life cycle compared to neutral and westerly intraseasonal periods (see Fig. 3). Anomalous moisture tendencies and TPW are reduced as well during easterly intraseasonal periods. For example, the moisture tendency maximum during easterly intraseasonal periods is weaker by 20% in advance of the NE quadrant during phase 7 when compared to the similarly placed maximum during neutral intraseasonal periods. The TPW anomalies during easterly intraseasonal periods are likewise reduced by approximately 30% in the SW quadrant of phase 3 compared to neutral intraseasonal periods (see Fig. 4). The sensitivity of anomalous convection to anomalous atmospheric moisture content is explored by comparing variations in diluted convective available potential energy (CAPE) anomalies for different intraseasonal periods at the end of the section.

Moistening of 1 and 0.5 mm day$^{-1}$, respectively, resulting from differences in vertical advection and precipitation occurs ahead of nascent and mature EWs during westerly and easterly intraseasonal periods (bottom rows of Figs. 8 and 9). Moistening by this term is stronger during westerly intraseasonal periods and not as consistent or robust throughout the wave’s life cycle.
during easterly intraseasonal periods when compared to neutral periods.

Similar to neutral intraseasonal periods, horizontal advection is responsible for most of the EW column-integrated moisture tendency during westerly and easterly intraseasonal periods (middle rows of Figs. 8 and 9). During westerly intraseasonal periods, total meridional advection contributes up to 40% more than the total zonal advection during the mature stages of the EW (not shown). During easterly intraseasonal periods, total meridional advection is up to 60% stronger than total meridional advection for the mature stages of the EW life cycle (not shown). Using Eq. (2) for westerly intraseasonal periods, total meridional advection is approximately 15% stronger during phase 3 than those during neutral intraseasonal periods. This result might be expected, given the stronger EW circulations during westerly intraseasonal periods (see Fig. 4; Maloney and Hartmann 2001; Rydbeck and Maloney 2014; Crosbie and Serra 2014). At the same phase and in the same location, the maximum for $-\left[\sigma^\prime q^\prime / \partial y\right]$ during easterly intraseasonal periods is 50% weaker (bottom row of Fig. 11).

The positive moisture tendency in the SW quadrant is reinforced by the mean zonal winds acting on the perturbation zonal moisture gradient ($-\left[\sigma^\prime q^\prime / \partial x\right]$) during both westerly and easterly intraseasonal periods (top rows of Figs. 10 and 11). For mature stages of the EW, this term contributes up to 2.5 mm day$^{-1}$ of the moisture tendency ahead of the wave during westerly intraseasonal periods and 3.5 mm day$^{-1}$ during easterly intraseasonal periods. Although not as strong, in the western region of the east Pacific warm pool, perturbation zonal winds acting on the mean zonal moisture gradient have a nonnegligible contribution to the column moistening on the poleward side of a mature EW.
up to 2 mm day$^{-1}$ during westerly intraseasonal periods (not shown).

The advection of mean moisture by the perturbation meridional winds is a dominant term in the moisture budget for EWs during neutral and westerly intraseasonal periods. The mean moisture gradient is represented in Fig. 7 by the mean TPW for each intraseasonal period. The mean 400-hPa pressure velocity is overlaid to indicate regions of deep rising motion indicative of the climatological ITCZ. Also, the tracks of the 700-hPa EW vortex centers are shown. Compared to neutral intraseasonal periods, the background TPW in the east Pacific warm pool during westerly intraseasonal periods is enhanced along the track of EWs, consistent with the findings of Crosbie and Serra (2014). TPW during easterly intraseasonal periods extends more continuously to the west and is narrower in latitude compared to neutral and westerly intraseasonal periods. Only slight variations in TPW occur near the Panama Bight among the different intraseasonal periods. Midlevel upward pressure velocities are also enhanced during westerly intraseasonal periods and suppressed during easterly intraseasonal periods compared to neutral intraseasonal periods. During all intraseasonal periods, EW tracks initially follow the equatorward side of the ITCZ. At these times, ahead of the wave, anomalous meridional winds are advecting the mean moisture equatorward. This process is particularly important when the anomalous meridional winds of the EW are strong (i.e., during westerly and neutral intraseasonal periods). Eventually, the tracks shift to the north, and the gradient of background meridional moisture on which the EWs operate is reduced, thus shutting down the moisture advection into the SW quadrant of the wave (see bottom rows of Figs. 6, 10, and 11).

To examine why comparable EW moisture tendencies during various background intraseasonal states result in different convection anomalies, we examine the diluted CAPE anomalies calculated using a prescribed entrainment rate of 0.18 hPa$^{-1}$ in the atmospheric boundary layer (below 950 hPa) and 0.002 hPa$^{-1}$ in the free

**Fig. 9.** As in Figs. 5 and 8, but for strong easterly waves during easterly intraseasonal periods.
troposphere. Sahany et al. (2012) showed these entrainment rates (referred to as entrainment case C2 in that paper) to produce a reasonable representation of tropical convective onset in the Pacific. The plume is assumed to be saturated at 1000 hPa. In using diluted CAPE, the effects of environmental mixing with a saturated plume are included. Figure 12 shows diluted CAPE anomalies for phases 3 and 7 of EWs during neutral (top row), westerly (middle row), and easterly (bottom row) intraseasonal periods. During phase 3, the nascent EW during westerly intraseasonal periods contains strong positive CAPE anomalies of 200 J kg\(^{-1}\) in the SW.
quadrant of the composite wave. This is double the anomalous diluted CAPE that occurs in a similar region of the EW during easterly intraseasonal periods. As a result, moisture anomalies during easterly intraseasonal periods are less supportive of convection anomalies during the easterly intraseasonal period. For phase 7, the mature wave during westerly periods has a local maximum of diluted CAPE located in advance of the NE quadrant that is ill defined at that phase for neutral and easterly intraseasonal periods. Diluted CAPE anomalies build more rapidly during westerly intraseasonal periods throughout the EW life cycle compared to other intraseasonal periods, providing a more favorable state for anomalous deep convection.

5. Adiabatic forcing

As noted earlier, moistening due to differences between vertical advection and precipitation suggests a possible role for large-scale forcing of convection in advance of EWs, especially if it can be shown that a component of low-level vertical velocity (and, hence, vertical moisture advection) is forced by the dynamics of the EWs in isolation from convection. Differences in vertical advection and precipitation are most notable in neutral and westerly intraseasonal periods. We focus on neutral intraseasonal periods because of the larger sample size. To explore the possible role of adiabatic forcing of vertical motion, a simple approximation for \( \mathbf{Q} \)-vector divergence is presented. Vertical motions induced by adiabatic forcing are likely amplified by diabatic processes in EWs such that the vertical velocity fields suggested by the \( \mathbf{Q} \)-vector divergence analysis should only be qualitatively compared to the observed variability. Kiladis et al. (2006) analyzed the \( \mathbf{Q} \)-vector divergence of African EWs in the southern wave track. They found that the location of vertical motion associated with adiabatic forcing quantified by the \( \mathbf{Q} \)-vector

![Fig. 12. The 2.5–12-day bandpass-filtered 700-hPa relative vorticity (\( \times 10^{-6}\) s\(^{-1}\); line contours) and diluted CAPE (J kg\(^{-1}\); color contours) anomalies composited for phases (left) 3 and (right) 7 of (top) neutral, (middle) westerly, and (bottom) easterly intraseasonal periods and strong easterly waves. The vorticity contour interval is \( 1 \times 10^{-6}\) s\(^{-1}\). Zero-vorticity contours are thick solid lines, and positive (negative) contours are solid (dashed). Phase and number of days included in each composite are shown on the right above each panel.](http://journals.ametsoc.org/jas/article-pdf/72/10/3850/3831945/jas-d-15-0056_1.pdf)
divergence was largely a result of the advection of perturbation vorticity by the mean thermal wind.

The approximate form of the \( Q \)-vector divergence utilized in Kiladis et al. (2006) but expanded in this study to include both meridional and zonal advection of perturbation vorticity by the mean thermal wind is

\[
\frac{pr}{R} \nabla_h \cdot \mathbf{Q}_h' = -\left( \frac{\partial T}{\partial y} \frac{\partial \xi_p'}{\partial x} + \frac{\partial T}{\partial x} \frac{\partial \xi_p'}{\partial y} \right),
\]  

where \( \mathbf{Q}_h' \) is the horizontal \( Q \) vector, \( T \) is the 10-day running mean temperature, \( \xi_p' \) is the deviation of the vertical vorticity from the 10-day running-mean vertical vorticity, \( p \) is the pressure, and \( R \) is the gas constant of dry air. The full \( Q \)-vector equation, as well as the motivation for using the approximate form, can be found in the appendix of Kiladis et al. (2006).

Similar to the results of Kiladis et al. (2006), the bulk of the adiabatic forcing calculated using Eq. (3) comes from the term representing the effect of perturbation vorticity advection by the mean thermal wind in the zonal direction, although the entire term will be shown. The \( Q \)-vector divergence at 900 hPa associated with strong EWs for select phases during neutral intraseasonal periods is shown in Fig. 13. Caution is taken when interpreting the results near the coast, as the assumptions for \( Q \) vectors may not be well constrained there, given the strong influence of the Sierra Madre. Away from the coast, \( Q \)-vector forcing associated with upward vertical motion is located ahead of maturing waves during phases 1, 3, 5, and 7. \( Q \)-vector forcing associated with downward motion is located behind the wave. Thus, adiabatic forcing of vertical motion is favored ahead of the wave, while descending motion is forced behind the wave. The horizontal temperature gradient vectors \(-[\partial T/\partial y, \partial T/\partial x]\) at 900 hPa show the advection of positive perturbation 900-hPa vorticity in advance of the wave during early and growing stages. The regions of \( Q \)-vector convergence at 900 hPa and moistening resulting from the difference of vertically integrated vertical advection and precipitation (Fig. 5) are located in very similar regions of the wave during early stages, indicating possible roles for moistening of the lower and middle troposphere by adiabatic forcing. While this analysis is suggestive, the total EW vertical motion anomaly field is complicated by influence from diabatic heating and is difficult to compare to the vertical velocity field implied by the \( Q \)-vector analysis. Consistent with this, the 2.5–12-day bandpass-filtered 900-hPa omega anomalies are nearly in phase with OLR anomalies (not shown).
6. Forcing of vorticity by convection

As can be seen from the EW composites in Figs. 1–3, east Pacific EWs begin as quasi-axisymmetric disturbances and evolve into waves with distinct SW–NE tilts. These horizontal tilts permit the wave to draw energy via barotropic conversion from the background cyclonic shear. The horizontal tilt of EWs and attendant barotropic conversions occur at both low- and midlevels (Rydebeck and Maloney 2014). We hypothesize that the anomalous convective centers in the SW and NE quadrants produce a favored SW–NE tilt by locally generating vorticity in those quadrants through vertical stretching. Vertical stretching has been shown to be important in the vorticity budget of EWs in the west Pacific (Yanai 1961; Lau and Lau 1992) and Caribbean (Yanai and Nitta 1967), particularly at low levels.

The vertically integrated EW vorticity budget can be expressed as

$$\frac{\partial \zeta_p}{\partial t} = -[\mathbf{v}_h \cdot \nabla_h (\zeta_p + f)] + \left[ (\zeta_p + f) \frac{\partial \omega}{\partial p} \right] + [\mathbf{v}_h \cdot (\zeta_h \omega)] + R,$$

(4)

where the local change of the vertical component of vorticity on the left-hand side results from horizontal vorticity advection, vorticity stretching/horizontal convergence of vertical vorticity, and effects of vertical motion on horizontal vortex lines, respectively, on the right-hand side. The residual $R$ includes processes that affect vorticity that are not represented in the other terms in Eq. (4), such as boundary layer friction, cumulus convection, other subgrid-scale processes, and any errors in the calculation of the terms. The vorticity budget is integrated from 1000 to 700 hPa to better capture the mechanisms modulating vorticity at low levels, specifically the low-level convergence induced by convection that may lead to vorticity stretching. All terms are anomalies obtained by applying a 2.5–12-day bandpass filter to the total terms.

In investigating the vorticity budget terms for neutral intraseasonal periods, we focus on the processes that perturb the initially quasi-axisymmetric vortex into one with a distinct SW–NE horizontal tilt and processes that help to maintain a tilted disturbance. Similar to previous studies on the vorticity budget of EWs in the Pacific and Caribbean (Yanai 1961; Yanai and Nitta 1967; Lau and Lau 1992), horizontal vorticity advection (top-right panel of Fig. 14) is the leading contributor to the overall tendency of EW vorticity anomalies (top-left panel of Fig. 14). The strong quadrature relationship of the vorticity tendency resulting from horizontal advection with the vorticity anomalies suggests that advection is largely responsible for the wave’s northwestward propagation but does not maintain or grow vorticity anomalies.

The contribution of the vorticity tendency not resulting from horizontal advection is shown in the bottom-left panel of Fig. 14. Processes included in this term are stretching, tilting, vertical advection, and processes not explicitly resolved in the budget. These processes maximize approximately 1/8 wavelength ahead of the EW and thus contribute both to the propagation and amplification of the wave. Grid-scale stretching of the vertical component of vorticity is shown in the bottom-right panel of Fig. 14. Unlike the horizontal advection of vorticity, the positive portions of the vertically integrated stretching term are largely in phase with the wave, directly support vorticity anomalies, and help to maintain a tilted wave structure. Stretching integrated below 700 hPa during phase 7 maximizes in the SW and NE quadrants of the wave and is greater than 0.018 s$^{-1}$ (12 h)$^{-1}$. The dotted solid line along the EW indicates the location of the vorticity stretching cross section averaged over a 1.5° width that is shown in Fig. 15.

Stretching in the mature wave of phase 7 is strongly preferred in low levels of the SW and NE quadrants (Fig. 15). Whereas stretching located in the center of the wave leads to a more axisymmetric growth of the vorticity, the stretching in the NE and SW peripheries of the wave supports a horizontal elongation of the wave. The stretching in phase 7 is collocated with regions of strong negative OLR anomalies representative of deep convection. We propose that local low-level vorticity stretching (which may also be referred to as local vorticity convergence) resulting from deep convection preferentially occurring in the SW and NE quadrants of EWs supports the observed horizontal tilt of the wave. The preferred horizontal tilt initializes when the wave is weak with the OLR maximum located in the SW quadrant (see phases 3 and 4 of Fig. 1), and the tilt intensifies as a complimentary OLR maximum develops in the NW quadrant (see phases 5 and 6 of Fig. 1). Such tilts are critical for the wave to derive energy from the background horizontal wind shear via barotropic conversion.

EW barotropic conversions and attendant tilts are known to occur at midlevels as well as low levels. The effects of vertical motion on horizontal vortex lines during phase 7, along the same cross section shown for vertical stretching, have local maxima exceeding $1.5 \times 10^{-6}$ s$^{-1}$ (12 h)$^{-1}$ in the middle and upper troposphere in both the SW and NE quadrants and thus result in EW horizontal tilts in regions above the strongest vorticity stretching (not shown). Last, the residual of the budget in the lower troposphere is overwhelmingly negative. The exclusion of boundary layer friction in the vorticity budget is likely the leading contributor to the residual
there. A budget residual maximum in the middle troposphere of $1.2 \times 10^{-6} \text{s}^{-1}$ suggests a possible role for unresolved processes, such as cumulus convective effects, for increasing midtropospheric vorticity (not shown).

### 7. Conclusions

EWs in the east Pacific have preferred areas of convection that evolve as the wave grows. Beginning as quasi-axisymmetric disturbances, the waves obtain strong SW–NE tilts by maturation. In this study, we hypothesize that preferred regions of EW convection located in the SW and NE quadrants of the wave force local vorticity stretching and elongate the wave horizontally in those directions.

Regions of anomalous convection are supported by collocated column-integrated moisture anomalies. While evaporation opposes the buildup of moisture early in the wave’s life cycle, horizontal advection overwhelmingly dominates the moisture budget of the waves. EWs propagating along the mean moisture gradient of the ITCZ advect moist air into the SW quadrant of the waves. The resulting column moisture anomaly maximum in that quadrant early in the wave’s life is supportive of coincident OLR anomalies. Later in the life cycle, as the wave moves northward, convection in mature waves is supported by moisture anomalies resulting from the advection of perturbation moisture anomalies by the mean zonal wind. At the same time, advection of mean moisture by the perturbation meridional winds in the SW quadrant declines as the wave moves away from the strong moisture gradient on the equatorward side of the ITCZ. In the moisture budget, vertical advection and precipitation largely cancel at the wave axis, but moistening by vertical advection exceeds drying by precipitation ahead of the wave. Behind the wave, precipitation exceeds vertical advection, resulting in drying. This signal is consistent with adiabatic forcing of rising motion ahead of the wave axis and sinking motion behind.

EWs are strongly modulated by the background fields through which they propagate. Previous studies have shown that EW kinetic energy and available potential energy are a strong function of the intraseasonal background state of the atmosphere (Maloney and Hartmann 2001; Aiyer and Molinari 2008; Rydbeck and Maloney 2014; Crosbie and Serra 2014). In this study, we suggest that intraseasonal variations of EW moisture that influence convective instability, as shown with a diluted CAPE calculation of a saturated plume, regulate the convective intensity of EWs. EW-diluted CAPE anomalies during
Easterly intraseasonal periods are reduced by half compared to westerly intraseasonal periods and, as a result, stifle EW convection anomalies. Regardless of intraseasonal phase, the spatial relationship between EW convection and circulation anomalies exhibits strong variations as a function of EW phase.

EW vorticity anomalies are propagated by horizontal advection and intensified by vertical stretching. The vertical stretching is coincident with regions of strong OLR anomalies, suggesting that deep convection modifies the vorticity structure of the EW: namely, an elongation of the wave’s circulation from SW to NE. The tilt of the wave from SW to NE is critical to the increase of EW kinetic energy by barotropic conversion cited in Rydbeck and Maloney (2014). EWs during easterly intraseasonal periods are, at times, not strongly convectively coupled and much slower to develop favorable horizontal tilts and corresponding EW kinetic energy increases.

More vigorous synoptic-scale vortices have been shown to favor enhanced tropical cyclogenesis in the east Pacific (Davis et al. 2008). Hence, in agreement with previous work, EWs during low-level westerly (easterly) periods that develop more (less) quickly are associated with enhanced (suppressed) periods of tropical cyclogenesis (Maloney and Hartmann 2000; Davis et al. 2008). The vast majority of perturbation available potential energy generation within east Pacific EWs and subsequent conversion to perturbation kinetic energy is associated with convective diabatic heating (e.g., Serra et al. 2010; Rydbeck and Maloney 2014). Because of this relationship, strongest considerations are typically placed on convective intensity when forecasting EW strengthening. The results presented herein suggest that a secondary and previously unidentified effect of convection is to tilt the wave in the horizontal, allowing the wave to draw energy from the background cyclonic wind shear. The tilt of the wave results from the distribution of convection within the envelope of EW vorticity. As a result, emphasis should be placed on the horizontal arrangement as well as on the intensity of convection within east Pacific EWs when considering EW growth patterns.

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