A 40-Year Climatology of Extratropical Transition in the Eastern North Pacific

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ABSTRACT

A 42-yr study of eastern North Pacific tropical cyclones (TCs) undergoing extratropical transition (ET) is presented using the Japanese 55-yr Reanalysis dataset. By using cyclone phase space (CPS) to differentiate those TCs that undergo ET from those that do not, it is found that only 9% of eastern North Pacific TCs that developed from 1971 to 2012 complete ET, compared with 40% in the North Atlantic.

Using a combination of CPS, empirical orthogonal function (EOF) analysis, and composite analysis, it is found that the evolution of ET in this basin differs from that observed in the North Atlantic and western North Pacific, possibly as a result of the rapidly decreasing sea surface temperatures north of the main genesis region. The presence of a strong, deep subtropical ridge extending westward from North America into the eastern North Pacific is a major factor inhibiting ET in this basin. Similar to other basins, eastern North Pacific ET generally occurs in conjunction with an approaching midlatitude trough, which helps to weaken the ridge and allow northward passage of the TC. The frequency of ET appears to increase during developing El Niño events but is not significantly affected by the Pacific decadal oscillation.

1. Introduction

Most eastern North Pacific tropical cyclones (TCs) develop over warm sea surface temperatures (SSTs) near the western coast of Central America and Mexico. The subsequent climatological track parallels the coastline before turning westward away from land, where these systems eventually dissipate over open ocean because of increasing vertical wind shear and cooling SSTs. The subtropical ridge situated over southwestern North America and extending over the eastern North Pacific for much of the hurricane season is largely responsible for this steering flow (e.g., Fig. 1). Some TCs move farther northward than average because of shifts in the subtropical ridge axis or weaknesses in the ridge induced by midlatitude troughs (Fig. 1b). These systems often bring rain to western Mexico and parts of the United States while exhibiting extratropical characteristics (Englehart and Douglas 2001; Ritchie et al. 2011; Wood and Ritchie 2013).

Extratropical transition (ET) can occur for this northward-moving subset of eastern North Pacific TCs.

The transition of a symmetric, warm-cored tropical cyclone into an asymmetric, cold-cored extratropical cyclone (e.g., Klein et al. 2000) is commonly observed in many of the ocean basins around the world, including the western South Pacific, the western North Pacific, the North Atlantic, and the southeastern Indian Oceans (Foley and Hanstrum 1994; Klein et al. 2000; Hart and Evans 2001; Sinclair 2002). Tropical cyclones that undergo ET can reintensify as extratropical cyclones and produce strong winds, high seas, and heavy rainfall in areas rarely affected by TCs (e.g., Jones et al. 2003). The timing of ET, as well as the strength of the post-ET cyclone, has been historically difficult to forecast.

When eastern North Pacific TCs move northward, they enter an environment characterized by decreasing SSTs, increasing vertical wind shear, and meridional air temperature and moisture gradients, all key ingredients for ET. However, it was noted by Jones et al. (2003) that the strong subtropical ridge in place for much of the eastern North Pacific hurricane season prevents ET from occurring. Despite this, individual cases of ET have been documented. For example, Tropical Storm Ignacio completed ET over open ocean during the 1997 eastern North Pacific hurricane season (Wood and Ritchie 2012). Ignacio was a minimal tropical storm that exhibited an unusual northward track and interacted with two separate midlatitude systems over the course of its
lifetime. After it completed ET west of California, it brought rain to the northwestern United States.

To evaluate the overall frequency of ET in this basin, investigate the behavior of the TCs that undergo ET, and add to the body of literature on ET in tropical basins around the world, this paper presents the first climatology of ET in the eastern North Pacific. It is constructed over the period 1971–2012 using the cyclone phase space (CPS; Hart 2003) methodology to define ET. This 42-yr period constrains the study to years with reliable satellite coverage. Section 2 describes the data used and the methodology applied in developing this climatology. Section 3 examines the characteristics associated with ET in the eastern North Pacific, including TC structural evolution, large-scale patterns, and variability. Section 4 compares cases undergoing ET with cases affecting southwestern North America, and section 5 offers concluding remarks and statements on future work.

2. Data and methodology

a. Data

The global 6-h spectral T159-model-resolution 40-yr European Centre for Medium-Range Weather Forecasts (ECMWF) Re-Analysis (ERA-40; Uppala et al. 2005) is used for the period 1971–78, and the global 6-h T255-resolution (nominally 0.7°) Interim ECMWF Re-Analysis (ERA-Interim; Dee et al. 2011) is used for the period 1979–2012. Global 6-h, 1° final operational global analysis (FNL) data from the Global Forecast System (GFS) are also examined for comparison with ERA-Interim over the period 2000–12. Global 6-h, 1° final operational global analysis (FNL) data from the Global Forecast System (GFS) are also examined for comparison with ERA-Interim over the period 2000–12. These data are obtained from the Research Data Archive (RDA), which is maintained by the Computational and Information Systems Laboratory (CISL) at the National Center for Atmospheric Research (NCAR) (at http://rda.ucar.edu). In late November 2013, the RDA made available a 1.25°-resolution version of the recently released T319L60 (~60 km) model-resolution Japanese 55-yr Reanalysis (JRA-55; Ebita et al. 2011) which covers the period 1958–2012 at 6-h temporal resolution. This dataset is compared to both ECMWF reanalyses as well as the FNL fields.

Optimum interpolation SST data at daily, 0.25° resolution (Reynolds et al. 2007) are obtained for the period 1982–2012 from National Oceanic and Atmospheric Administration/Earth System Research Laboratory/Physical Sciences Division (NOAA/ESRL/PSD; at

![Fig. 1. Average 500-hPa geopotential height (m) from the JRA-55 over 1971–2012 and tracks of TCs that exist for the majority of their lifetimes during the months of (a) June and (b) September.](image)
Because of its greater period of coverage and its use in generating the JRA-55 dataset, monthly 1° Centennial in situ Observation-Based Estimate (COBE; Ishii et al. 2005) SST fields are also examined. Position and intensity data every 6 h are obtained from the National Hurricane Center best track [Hurricane Database (HURDAT); Davis et al. 1984] over the 42-yr period, resulting in a database of 631 TCs.

b. Analysis methods

The structural evolution of each of the 631 TCs is described using CPS (Hart 2003), a method that evaluates the thermal characteristics of the storm using a measure of geopotential thickness asymmetry (the $B$ parameter) and the thermal wind. The $B$ parameter is computed by subtracting the average 900–600-hPa geopotential thickness on the right side of the storm track from that on the left side of the storm track; low values represent thermally symmetric systems, while high values represent thermally asymmetric systems. The thermal wind,

$$-V_T = \frac{\partial (\Delta Z)}{\partial \ln p},$$

is computed by linear regression of the difference in maximum and minimum height between isobaric levels within a radius of 500 km, $\Delta Z$. In CPS, the thermal wind is evaluated within two different layers: 900–600 hPa (lower level) and 600–300 hPa (upper level). Positive values of $-V_T$ mean the geostrophic wind decreases with height in the layer (warm-cored tropical systems), while negative values of $-V_T$ mean the geostrophic wind increases with height in the layer (cold-cored extratropical systems).

The phases are separated by a $B$ parameter value of 10 m and a thermal wind ($-V_T$) value of 0 m s$^{-1}$ (Fig. 2). Tropical systems have symmetric structures ($B < 10$ m) and warm cores ($-V_T > 0$ m s$^{-1}$). Asymmetric storms have large $B$ values ($> 10$ m), and a cold core is depicted by a negative thermal wind value ($-V_T < 0$ m s$^{-1}$). As a result, TCs undergoing ET have a $B$ value that increases to greater than 10 m and $-V_T$ values that change sign from positive to negative. To exclude weak transition cases in this study, TCs had to achieve a $B$ value of at least 11 m and a $-V_T$ value of $-1$ m s$^{-1}$ or less to qualify as ET cases. Only the subset of TCs that undergo ET are retained for further analysis. The time at which ET occurs for this subset is then defined as the first synoptic time the TC has a $B$ value greater than 10 m and a $-V_T$ value less than 0 m s$^{-1}$.

Grid Analysis and Display System (GrADS) scripts provided by Dr. Hart (http://moe.met.fsu.edu/~rhart/phasescripts/) were utilized to facilitate the calculation of CPS parameters from the reanalyses.

Empirical orthogonal function (EOF) analysis (e.g., Bretherton et al. 1992) in the form of singular value
decomposition (SVD) is applied to objectively evaluate the dominant large-scale patterns associated with ET in the eastern North Pacific. Similar to previous work (Wood and Ritchie 2013), storm-centered fields are first extracted from the reanalyses. The mean of the extracted fields is used to compute the anomalies on which EOF analysis is then performed. The statistical significance of the resulting patterns are evaluated using the North test (North et al. 1982) and separately considered for physical significance.

c. ERA-Interim versus FNL

ERA-Interim and ERA-40 were initially chosen for use in this study because of the temporal coverage of the data as well as the generally positive performance of the ECMWF models with respect to TCs (e.g., ECMWF 2007). As FNL data are regularly used to drive regional models, including numerical simulations of TCs, these data are also utilized in this study for comparison with ERA-Interim.

Hart (2003) employed a phase space constructed with the $B$ parameter and the 900–600-hPa thermal wind to characterize the structural evolution of a TC. However, comparing the ERA-Interim and FNL values in this phase space reveals discrepancies in the evaluation of the lower-level thermal wind between the two datasets, while the upper-level (600–300 hPa) thermal wind values appear to be more similar. Hurricane Hilary (2011), a TC that peaked at category 4 intensity, exemplifies this discrepancy (Fig. 3). The 900–600-hPa thermal wind (Fig. 3a) from the FNL depicts a warm core for almost all of Hilary’s lifetime, but the same parameter calculated from the ERA-Interim indicates a cold-core system despite the hurricane intensity of the TC during this time. Conversely, the 600–300-hPa thermal wind values remain positive for about the same period of time in both datasets (Fig. 3b), though ERA-Interim depicts a weaker warm core.

These differences inspired the addition of a third dataset, the recently released JRA-55, to the comparison. Unlike the ECMWF reanalyses, JRA-55 includes wind profile retrievals for TCs in order to improve their representation in the data and follows the same approach used in the JRA-25 (Hatsushika et al. 2006). Their study showed that this inclusion improves TC representation in the JRA-25 in observation-sparse areas such as the eastern North Pacific. Much like the JRA-25 (not shown), the JRA-55 does depict...
a warm core in the 900–600-hPa layer for Hurricane Hilary during the same period as the FNL, albeit a weaker one (Fig. 3a), and the 600–300-hPa thermal wind values are similar to the ERA-Interim results (Fig. 3b).

Figure 3c shows the zonal anomalies of geopotential height from the ERA-Interim, FNL and JRA-55 datasets for Hurricane Hilary at 0000 UTC 25 September when the TC had an intensity of 115 kt. At this time the TC exhibits warm-core structure in the lower levels (900–600 hPa) in both the FNL and JRA-55 but cold-core structure in the ERA-Interim. Conversely, the upper-level structure depicts a warm core in all three datasets. Though the warm core is far stronger in the FNL data, both FNL and JRA-55 depict TC structure similar to that found in the literature (e.g., Hart 2003). However, the ERA-Interim does not, centering the maximum warm anomaly in the middle of the troposphere rather than near the surface.

To expand the scope of the comparison, CPS parameters are calculated for all eastern North Pacific TCs over the 2000–12 hurricane seasons from the ERA-Interim, JRA-55, and FNL datasets and shown in Figs. 4 and 5 as a set of frequency diagrams. The ERA-Interim results produce a frequency peak that corresponds to a high number of cold-core occurrences in the 900–600-hPa layer, while the FNL and JRA-55 exhibit a strong warm-core peak frequency (Figs. 4a,c,e). Conversely, the results from both datasets using the 600–300-hPa thermal wind resemble each other more closely, with a slightly stronger warm-core frequency peak from ERA-Interim compared to FNL (Figs. 4b,d). The JRA-55 showcases the strongest frequency peak as well as a narrower range of $B$ values compared with the other datasets (Fig. 4f). The frequency diagrams in Fig. 5 provide supporting evidence of the ERA-Interim bias toward cold-core values in the 900–600-hPa layer compared with the FNL and JRA-55 results.

Zonal geopotential height anomalies such as those presented in Fig. 3c reveal that many eastern North Pacific TCs have a warm core centered aloft in the ERA-Interim fields (not shown). On the other hand, both the JRA-55 and the FNL data generally depict a peak anomaly near the surface much like that shown using Navy Operational Global Atmospheric Prediction System (NOGAPS) data in Hart (2003). As a result, the analysis presented in Wood (2012) using ERA-40 and ERA-Interim data as well as a modified phase space of the 600–300-hPa thermal wind and the $B$ parameter has been updated to use the JRA-55 dataset and the original CPS methodology from Hart (2003). Many but not all TCs that completed ET according to ERA-Interim also completed ET in the JRA-55, and only those that completed ET in JRA-55 will be discussed in the following sections. All atmospheric fields presented in the following sections are derived from the JRA-55 dataset.

It is somewhat surprising to discover significant differences between the results of the same methodology calculated using two different datasets, as is the case for Hurricane Hilary (Fig. 3). Note that the data depicted in Figs. 4 and 5 are calculated for all eastern North Pacific
TCs during 2000–12, not just ET cases. However, analysis of non-ET cases in the eastern North Pacific is beyond the scope of this paper, and a manuscript that further examines the structural evolution of all TCs in this basin and compares CPS results from seven reanalysis datasets over the 1979–2010 period is in preparation.

3. Characteristics of extratropical transition in the eastern North Pacific

Over the period 1971–2012, 55 of the 631 named TCs underwent ET according to the JRA-55-based phase space described by the \( B \) parameter and the 900–600-hPa thermal wind, composing about 8.7% of the dataset (Table 1). Most of these ET events (65.5%) occur in September and October, when the subtropical ridge is weaker and midlatitude troughs protrude farther south (e.g., Fig. 1b). As a result, the peak in ET frequency is shifted to later in the season compared with the peak genesis frequency (Fig. 6). Figure 7 shows the track density of ET cases separated by the month in which ET occurred as well as the average 500-hPa geopotential height at ET time. The infrequent development of TCs in November is partially responsible for the dearth of ET events during this month. Tracks of TCs undergoing ET are dominantly steered northwestward around the subtropical ridge and tend to cluster near the coast of North America. The greatest spread occurs in August and September (Figs. 7d,e), when strong easterly steering flow can cause TCs to move far westward before finally moving around the periphery of the ridge or interacting with a midlatitude trough.

Overall, the frequency of ET in the eastern North Pacific is far lower than that found in other basins. In the North Atlantic, 46% of TCs complete ET (Hart and Evans 2001). The western North Pacific boasts the largest number of ET events, with 274 TCs completing ET over the 1979–2004 period (Kitabatake 2011), and about one-third of TCs in the southwest Pacific undergo ET (Sinclair 2002).

a. Structural evolution

Most eastern North Pacific TCs remain symmetric throughout their lifetimes. They develop as symmetric \((<10\,\text{m})\) or, less often, asymmetric \((>10\,\text{m})\)
warm-core systems before dissipating or transitioning to a cold-core structure (Fig. 8a), and those that become cold core tend to remain symmetric in the process (Fig. 8b). The subset of TCs that become extratropical follow a similar trend by first losing their warm-core structure before becoming asymmetric as they complete ET (Fig. 8c). This stands in contrast to the process of ET in other basins, such as the North Atlantic, where TCs become asymmetric while maintaining a warm core before becoming extratropical (Hart 2003).

One factor that may influence this tendency to retain symmetry is the sharp SST gradient that exists to the northwest of the genesis region in the eastern North Pacific basin, where SSTs decrease rapidly with increasing latitude and the 26°C isotherm often lies between 15° and 20°N (not shown). Figure 9 shows a subset of 29 ET cases that remained over ocean for 24 h after completing ET over the 1982–2012 period. The average SSTs near the TC center decrease from about 25°C 48 h before ET to nearly 21°C 24 h after ET. There is a corresponding weakening trend as shown by the average mean sea level pressure (MSLP) near the storm center during this time frame. These cool SSTs inhibit TC development and may contribute to the erosion of the system’s warm core at lower levels in the troposphere before midlatitude westerlies begin to induce asymmetries in the TC’s thermal structure (e.g., Ritchie and Elsberry 2001). Other contributing environmental factors include increasing vertical
wind shear, decreasing atmospheric moisture, and greater atmospheric stability.

Furthermore, this sharp SST gradient stands in contrast to the warmer waters found in the North Atlantic and western North Pacific, which may help explain the different trajectory eastern North Pacific ET events take through phase space (Fig. 10). Phase space frequency diagrams comprised of $-V_T^L$ and $B$ are calculated from the JRA-55 and normalized by the number of TCs in the 2001–10 period (Figs. 10a–c). North Atlantic and western North Pacific TCs generally exhibit deeper low-level warm cores than those found in the eastern North Pacific, and transitioning cases in both basins tend to develop thermal asymmetries while maintaining a warm core prior to completing ET. Note the tendency (Figs. 10e,f) for the TC trajectories to cross the $B = 10$ m line while still maintaining a warm core and then becoming cold core. Conversely, eastern North Pacific TCs generally maintain their symmetry while changing from warm to cold core before becoming extratropical (Fig. 10d). Overall, ET itself tends to be weaker for eastern North Pacific TCs, as the subsequent asymmetry of the extratropical system is much greater in the North Atlantic and western North Pacific (Figs. 10a–c), and the strength of the subsequent cold core tends to be larger as well. Part of the explanation for these differences may relate to the other two basins lying off the eastern side of a continent over which midlatitude troughs can intensify before they propagate over the ocean. Unlike the North Atlantic and western North Pacific, where TCs move around the periphery of a subtropical ridge, eastern North Pacific TCs must move through it, which may also play a role in the generally weaker ET in this basin.

Figure 11a shows the composite evolution of JRA-55 isentropic potential vorticity (IPV) within a 500-km radius from the storm center over a 96-h period centered on ET time. This field is calculated from a subset of 35 cases comprised of those TCs that could still be tracked for 48 h after ET time, as almost all of the 55 transitioning TCs underwent ET less than 48 h prior to the end of the best track and most within 24 h of the end of the record. The composite TC circulation anomaly during the 48 h leading up to ET is weak, though it intensifies after ET time. Upper-level IPV slowly increases until ET occurs and increases more rapidly after this point. This implies that the midlatitude trough is still

![Figure 8](image-url)  
**Fig. 8.** CPS frequency diagrams for (a) all 631 TCs, (b) 576 TCs that did not undergo ET, and (c) 55 TCs that completed ET. The black arrows denote the progression of the average TC through phase space during its lifetime.

![Figure 9](image-url)  
**Fig. 9.** NOAA optimum interpolation (OI) SST (°C; blue) and JRA-55 mean sea level pressure (hPa; green) from a $2^\circ \times 2^\circ$ box average centered on each TC relative to ET time. Thick line represents average value; thin line represents first standard deviation. The vertical dashed line indicates ET time. A subset of 29 TCs from 1982 to 2012 is included in this analysis.
approaching the TC circulation at the time of ET as found by using CPS. The large standard deviation values highlight the variability in the individual IPV fields of the TCs included in the analysis.

Following the methodology of Kofron et al. (2010), IPV anomalies are computed by subtracting the \(-48\)-h upstream trough IPV from the total IPV field (Fig. 11b). The upper-level IPV signature from the 35-case subset resembles that found for post-transition dissipation cases (Kofron et al. 2010), yet the persistent low-level IPV through 48 h implies that some cases do reintensify after ET is complete. For seven cases that reintensify in the IPV fields after ET, it becomes apparent that strong midlatitude flow is involved in the ET process (Fig. 11c). Composites of geopotential height for these cases reveal that the TC is stronger initially and that an incoming midlatitude trough induces a break in the subtropical ridge before interacting with the TC (not shown). The pattern of standard deviations from the 35-TC subset (Fig. 11b) resembles the overall pattern for the re-intensifying ET cases, so these strong cases explain much of the variability found in the IPV fields.
**b. Large-scale patterns**

As mentioned in section 1, eastern North Pacific TCs have largely westward tracks as a result of the steering flow induced by the subtropical ridge, but some cases do move northward. These less common tracks often occur when the ridge has weakened or is broken by a mid-latitude trough passing to the north or when the TC itself is strong enough to move along the planetary vorticity gradient. As these systems move poleward, they approach the baroclinic zone and thus increase the probability of ET (e.g., Fig. 12).

In addition to the sharp negative SST gradient, the area northwest of the main genesis region is characterized by increasing vertical wind shear and decreasing atmospheric humidity (Figs. 12a–d). The subset of eastern North Pacific TCs that move poleward when conditions are favorable (Figs. 12e,f) thus encounter three environmental factors known to be detrimental to TC development almost simultaneously, and many storms rapidly weaken and dissipate shortly after entering this region. As a result, few TCs survive long enough in these conditions to reach the baroclinic zone and begin the ET process.

To examine the large-scale patterns that are conducive to ET in the eastern North Pacific, objective characterization of 500-hPa geopotential height patterns relative to ET time is accomplished using SVD analysis (section 2). Three separate EOFs are computed: 24 h prior to ET time, ET time itself, and 24 h after ET time. As discussed in section 2b, ET time is defined as the synoptic time period just after the system entered the asymmetric cold-core quadrant ($-\frac{\nabla L}{\nabla T} < 0 \text{ m s}^{-1}; B > 10 \text{ m})$ in CPS.

The first two EOFs from this analysis are shown in Fig. 13. The first EOF explains 47%, 56%, and 55% of the variance of the fields at each time, respectively, and the second EOF explains 20%, 16%, and 17%, respectively. The positive mode of EOF1 depicts a pattern of ET that is quite similar to the classic pattern found in other basins. The TC moves through a break in the subtropical ridge situated over Mexico and gradually weakens before interacting with a deep midlatitude trough and merging with it (Figs. 13a–c). Conversely, the negative mode of EOF1 exhibits a much stronger, higher-amplitude ridge and a much weaker trough well to the north of the TC moving through a weakness in the ridge. This pattern of evolution represents weaker TCs that interact with a weak midlatitude trough and dissipate shortly after becoming extratropical (Figs. 13d–f).

The positive and negative modes of the second EOF demonstrate two other trough interaction patterns. The first EOF2 mode is similar to the positive EOF1 mode (Figs. 13g–i) but with an initially weaker TC 24 h prior to ET time; a stronger, higher-amplitude ridge; and a slightly weaker midlatitude trough. Note that the EOF pattern for 24 h before ET is flipped, with the negative mode developing into the positive mode for ET time and 24 h after ET and vice versa. The other EOF2 mode depicts a TC of similar intensity to the positive mode of EOF1 which moves northward through a break in the ridge. However, the incoming midlatitude trough is weaker, more negatively tilted, narrower, and overall
more suggestive of a cutoff low at ET time as it slowly merges with the TC. All four patterns do emphasize the common characteristics of a weakness or break in the subtropical ridge and an equatorward-protruding mid-latitude trough when ET occurs in the eastern North Pacific.

Storm-centered composite fields well characterize the average evolution throughout the troposphere as an eastern North Pacific TC undergoes ET (Fig. 14). At 24 h before ET, the composite TC exhibits a closed circulation from the surface through 500 hPa as it moves northward through a weakness in the deep, strong subtropical ridge. The approaching midlatitude trough begins to merge with the TC at ET time; the TC circulation has degenerated into an open wave at 500 hPa, though it remains closed at the surface and 700 hPa. Finally, 24 h after ET, the TC has been fully captured by the now deepened midlatitude trough. The deep subtropical ridge has weakened throughout the troposphere and retreated eastward, though it begins to redevelop westward on the equatorward side of the extratropical cyclone by 24 h after ET at 700 hPa (Fig. 14i).

These composite fields also highlight the relatively symmetric structure of the TC throughout the ET process. The circulation center at 700 and 500 hPa remains relatively collocated with the minimum MSLP, and the overall strength continues to decrease at each level. Though specific cases do reintensify post-ET, the general trend is for eastern North Pacific TCs to dissipate after the completion of ET. Both aforementioned characteristics differ from the trends observed in the North Atlantic and western North Pacific, as reintensification occurs with greater frequency in both basins and increasing asymmetry is generally observed for transitioning cases (e.g., Figs. 10d–f).

c. Interannual variability

There is little interannual variability in the number of ET events that occur in the eastern North Pacific during
FIG. 13. The first two EOF patterns for storm-centered 500-hPa geopotential height (m) 24 h before ET, at ET time, and 24 h after ET. The black dots indicate the location of the TC. EOF1 explains 47%, 56%, and 55% of the variance, respectively. EOF2 explains 20%, 16%, and 17% of the variance, respectively. Note that the second EOF for 24 h before ET is flipped. Both EOFs are statistically significant by the North test.
the period 1971–2012 with a mean of 1.3, median of 1, and standard deviation of 1.26 events each year. However, there is a suggestion that the El Niño–Southern Oscillation (ENSO) may have an impact on eastern North Pacific ET frequency through changing SSTS and the associated atmospheric circulation patterns. The most ET events in a given year (6) occurred in 1983, and the second most ET events (4) occurred in 2009. The 1983 eastern North Pacific hurricane season took place as a strong El Niño event waned; the Niño-3 3-month average SST anomalies (compared to the 1981–2010 average) peaked near 2.8°C during the December–February period prior to the start of the hurricane season. The 2009 hurricane season took place as an El Niño event was developing, and during this time the Niño-3 July–September anomaly was 0.8°C.

Eleven of the years from 1971 to 2012 (26%) did not produce any ET events. Eight of these 11 years (1973, 1979, 1988, 1995, 1996, 1998, 2007, and 2010) had seasonal activity below the 42-yr average of 15 TCs. In fact, 2010 is one of the least active seasons on record with 7 named TCs (Stewart and Cangialosi 2012). In addition, the seasons of 1973, 1974, 1988, 1998, 2007, and 2010 occurred during developing La Niña events. The remaining two years, 2005 (15 TCs) and 2008 (16 TCs), featured normal or near-normal activity and no or weakly negative SST anomalies in the Niño-3 region for much of the season.
To quantify the potential impact of SST anomalies on ET frequency, annual and monthly ET counts are correlated with 3-month average Niño-3 anomalies (Table 2). All statistically significant correlations at the 95% confidence level are positive. For yearly ET counts, average Niño-3 anomalies for April–June through October–December are significantly correlated. For monthly ET counts, only August and September exhibit any statistically significant correlations with 3-month Niño-3 averages, and the statistically significant relationships in September are restricted to the latter part of the hurricane season. Of the four years with Niño-3 average anomalies below \(2^{\circ}\)C during the June–October period (1973, 1988, 2007, and 2010), the months in which ET has been observed, no ET events occurred. Conversely, all of the years with June–October Niño-3 average anomalies above \(1^{\circ}\)C (1972, 1976, 1982, 1987, and 1997) produced ET events, and all but 1987 produced more than one.

The distribution of average TC genesis and ET frequency between the 5 El Niño years, 4 La Niña years, and remaining 33 neutral years are also shown in Table 2. The composite monthly SST and 500-hPa geopotential height anomalies as well as the tracks of TCs undergoing ET during the warm years are shown in Fig. 15. These results imply that El Niño events developing during the hurricane season may increase the likelihood of ET. Warmer SSTs may extend TC lifetimes and
thus increase the chance of a TC approaching the baroclinic zone and beginning the ET process. The associated changes in the large-scale flow may shift the baroclinic zone southward and thus decrease its distance from the genesis region. More frequent breaks in the subtropical ridge, as implied by the below average 500-hPa geopotential heights along the coast of North America during warm years (Fig. 15a), may steer TCs farther northward than normal. Conversely, cooler SSTs associated with La Niña events may reduce TC lifetimes and result in fewer TCs entering the baroclinic zone, and the baroclinic zone itself may retreat northward. In addition, fewer breaks in the subtropical ridge or an overall stronger ridge, as implied by the above average 500-hPa geopotential heights north of 20°N (Fig. 15b), could steer TCs westward more frequently.

Given the 42-yr coverage of this study, some investigation of the influence of the Pacific decadal oscillation (PDO; Mantua et al. 1997) can be accomplished. A 5-month running mean is calculated from the monthly PDO index (http://jisao.washington.edu/pdo/) and compared with annual and monthly ET frequency. However, no significant correlations are found between annual ET counts nor monthly ET counts and the 5-month means.

Figure 16 shows the average SST anomalies over the June–October period (5-month average centered on August) for six years with an average PDO index greater than 1 and eight years with an average PDO index less than −1. The warm composite averaged 2.3 ET events per year, while the cool composite averaged 1.1 ET events per year. This is not a significant difference at the 95% level. Though the 500-hPa geopotential height anomalies are slightly below average during warm years and above average during cool years, this implies that the PDO does not significantly affect the likelihood of ET. Instead, the phase of ENSO appears to have a stronger impact on ET frequency.

4. Effects of ET on North America

As previously shown (e.g., Fig. 7), TCs undergoing ET often make landfall along the western coast of North America. Wood and Ritchie (2013) investigated 167 TCs that affected southwestern North America over the period 1989–2009. Many of the cases that brought rainfall to the southwestern United States interacted directly with a midlatitude trough or moved through a weakness in the subtropical ridge induced by a trough. These TCs subsequently recurved and moved over land. This process often resembled ET, which behooves a comparison between that dataset and the 55 TCs identified as completing ET in this study.
During the years 1989–2009, 23 TCs completed ET: 20 of which coincide with the set of TCs affecting southwestern North America explored by Wood and Ritchie (2013) (Fig. 17). Over this period, ET cases impact Mexico and/or the southwestern United States every month from June through October, with the highest frequency occurring in September (Fig. 17a). Three ET cases in 1990 and two each in 1989, 1997, and 2009 brought rain to parts of Mexico and/or the United States, while no ET cases occurred in 1995, 1996, 1998, 2005, 2007, or 2008 (Fig. 17b).

Both EOF1 and EOF2 (Fig. 13) are represented by this subset of TCs, implying that no single statistically significant pattern is responsible for the majority of ET cases impacting North America. The average rainfall pattern from the 20-TC subset (Fig. 18) peaks along the western coast of Mexico and spreads into the southwestern and central United States as the extratropical system moves northeastward. Much of the heaviest precipitation in Mexico is driven by topographic influences.

5. Summary and conclusions

This study presents a 42-yr climatology of extratropical transition in the eastern North Pacific using cyclone phase space parameters calculated from the JRA-55 dataset to characterize ET. The process of ET occurs less frequently in the eastern North Pacific compared with other basins in which TCs develop in part because of the presence of a strong and deep subtropical ridge during much of the hurricane season. However, 9%, or 55, of the 631 TCs during the 1971–2012 period are shown to complete ET. In the eastern North Pacific, transitioning cases tend to quickly move onshore, often leading to impacts in Mexico and the southwestern United States.

The process of ET in the eastern North Pacific appears to differ from that observed in the North Atlantic and western North Pacific. In the two latter basins, the TC generally maintains its warm core while becoming increasingly asymmetric before completing ET. Conversely, in the eastern North Pacific, storms undergoing ET tend to retain their symmetry but lose the warm core before becoming extratropical. The prevalence of cool SSTs to the west and north of the main development region in the eastern North Pacific likely contributes to this pattern of structural evolution.

Objective analysis and storm-centered composites reveal that, similar to other basins, ET in the eastern North Pacific tends to occur in conjunction with the approach of a midlatitude trough. However, the presence of a strong, deep subtropical ridge extending westward into the eastern North Pacific has a significant impact on the reduced frequency of ET compared with other basins. In the eastern North Pacific, ET rarely occurs without the presence of a strong midlatitude trough that can induce a weakness or break in the subtropical ridge, allowing the TC to move far enough poleward to interact with the trough. Many TCs that do propagate northward during an interaction with a midlatitude trough impact North America during or after completing ET.

The patterns of SST in the eastern North Pacific appear to have some influence on ET frequency. This may occur directly as SSTs influence both TC lifetimes and the tracks they can take (moving more poleward when SSTs are warmer at higher latitudes) and indirectly because of shifts in the large-scale flow associated with ENSO. Over the period of study, ET events occur with greater frequency during El Niño years than neutral years and do not occur at all during La Niña years. The PDO does not appear to have a significant influence on ET frequency in the eastern North Pacific.

Ongoing work includes further investigation of the differences in TC structure representation between reanalyses due to the discrepancies found between CPS results in the ERA-Interim and the JRA-55 datasets.

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