Comparison of Wave Packets associated with Extratropical Transition and Winter Cyclones

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Developing wave packets associated with the extratropical transition (ET) of tropical cyclones and winter cyclones in the Western North Pacific (WNP) and Atlantic basins are diagnosed observationally by compositing reanalysis data over a 32 year period. While the development of winter cyclones amplifies a weaker wave packet moving through the mid-latitude storm track, there is no indication of an upstream disturbance during ET; thus, on average, the wave packet is generated by the ET process. In the WNP, ET and winter cyclone wave packets have comparable group velocity and amplitude relative to climatology, whereas ET wave packets have relatively longer wavelength and near-zero group velocity during the ET process; ET events also have a detectable signal further downstream. Wave packets associated with winter cyclones in the Atlantic basin have greater amplitude and have a detectable signal further downstream relative to those associated with ET. Near the surface, winter cyclones are characterized by larger meridional heat fluxes relative to ET. WNP ET cyclones are characterized by larger meridional moisture flux convergence and thus latent heat release relative to their winter counterparts, while Atlantic basin ET and winter cyclones have similar moisture flux convergence. This result suggests that the wave packets associated with ET cyclones are related to diabatic processes and could explain why the amplitude of Atlantic basin ET wave packets are smaller than winter cyclones. Finally, the greater baroclinicity during winter does not seem to influence the downstream packet amplitude in either basin.
1. Introduction

The extratropical transition (ET) of tropical cyclones (TCs) is often associated with
the amplification of an upper-tropospheric anticyclone downstream of the TC (e.g., Jones
et al. 2003; Agusti-Panareda et al. 2004; Harr and Dea 2009). The development of this ridge
occurs in response to a number of dynamical mechanisms, including the adiabatic interaction
of the TC circulation with the midlatitude waveguide (e.g., Ferreira and Schubert 1999;
Riemer et al. 2008), and the diabatic outflow from either the TC or the baroclinic zone that
often develops on the down-shear side of the cyclone (e.g., Bosart and Dean 1991; Harr and
Elsberry 2000; Riemer et al. 2008; Riemer and Jones 2010; Torn 2010). The development of
this downstream ridge serves as an impulsive disturbance on the midlatitude flow, which can
give rise to wave packets and downstream development (e.g., Simmons and Hoskins 1979;
Chang and Orlanski 1993; Orlanski and Sheldon 1995; Hakim 2003), spreading the impact
of the ET event further downstream from the cyclone itself.

Previous studies have found that the downstream response of the midlatitudes to ET
varies from case to case, depending on the phasing of the TC with midlatitude features,
and that downstream development does not require a reintensifying tropical cyclone (e.g.,
Harr and Dea 2009; Riemer and Jones 2010). These wave packets are often associated with
forecast errors well downstream of the ET (e.g., Jones et al. 2003; Harr et al. 2008; Anwender
et al. 2008). Consequently, this motivates a deeper understanding of the development and
propagation of these wave packets over many cases.

ET events are not the only phenomenon that can create impulsive wave packets within
the midlatitude flow. Much of the literature on this topic has focused on the role of mid-
latitude cyclones. Both observational (e.g., Orlanski and Katzfey 1991; Chang 2000; Hakim
2003) and statistical (e.g., Chang 1993, 1999; Chang and Yu 1999) studies suggest that
these wave packets are energy sources whereby upstream disturbances seed downstream dis-
turbances. Moreover, midlatitude forecast errors develop and propagate similarly to wave
packets (e.g., Hakim 2005). For example, forecasts have also been shown to be sensitive to
wave packet initialization (e.g., Langland et al. 2002) and that the impact of assimilating targeted observations spreads downstream as a wave packet (e.g., Szunyogh et al. 2000).

The goal of this work is to compare wave packets associated with ET with those associated with winter cyclones in the Western North Pacific (WNP) and Atlantic basins. In particular, this study evaluates whether one can reject the null hypothesis that there is no meaningful difference in the genesis, structure, and propagation of wave packets associated with ET and winter cyclones. This hypothesis is tested by comparing a large sample of wave packets associated with ET and winter cyclones by averaging over many cases and applying the packet diagnostic technique outlined in Hakim (2003) to quantitatively compare packet properties. While previous work on the downstream impact of ET has focused on individual case studies or a small number of cases, this study bridges the ET and wintertime wave packet literature for a large sample.

The remainder of the paper proceeds as follows. Section 2 describes the dataset and methods used to compute the wave packets and their properties. Results of the calculations are presented in section 3 followed by a summary and conclusions in section 4.

2. Method

Wave packets associated with ET and winter storms in the Western North Pacific (WNP) and Atlantic basins are evaluated by compositing atmospheric fields, similar to the strategy employed by Hakim (2003). These two basins are chosen because of the overlap between the region of maximum ET and winter cyclogenesis frequency (e.g., Sanders and Gyakum 1980; Klein et al. 2000; Hart and Evans 2001) and the numerous studies of ET in each basin.

Atmospheric fields are taken from version 1 of the National Centers for Environmental Prediction (NCEP) Climate Forecast System (CFS) Reanalysis dataset (Saha et al. 2010) on mandatory constant pressure surfaces, with horizontal and temporal resolution of 2.5°
and 6 h, respectively, from 1979-2010. Anomalies are defined as deviations from a moving-average climatology, meaning that every day of the year has a unique climatology, which is defined as the average of all daily fields within ± 14 d of the day of interest during the 32 year period. This method of computing the climatology has the advantage of producing a smooth climatology from one day to the next. Statistically significant anomalies are defined in terms of a two-side Student’s t test with a threshold of 95%.

The sample of winter cyclone wave packets is determined by identifying rapidly deepening cyclones at the location of highest frequency in each basin. As in Hakim (2003), baroclinic cyclones are defined as local maxima in 1000 hPa geostrophic relative vorticity exceeding $10^{-4}$ s$^{-1}$ during November–March. Several additional checks are employed to ensure that individual cyclone events are identified. Any cyclone that is within 25° of another cyclone is removed from consideration. To ensure that the same event is not identified twice, any cyclone within 15° latitude and longitude of a previously identified cyclone during the previous 24 h is removed from the list of potential candidates. Hereafter, “t = 0” refers to the time when the cyclone exceeds the above critical value, which approximates when the cyclone reaches the mature stage. Based on these criteria, the maximum number of cyclone events in the WNP occurs within 35°N-40°N, 145°N-155°N, while in the Atlantic the maximum number occurs within 40°N-45°N, 55°-65°W; the latter region is slightly to the east of the box used in Hakim (2003). These boxes contain 281 and 334 cases in the WNP and Atlantic Basins, respectively.

Wave packets for ET cases are determined by evaluating all TCs contained in the Joint Typhoon Warning Center (JTWC) WNP best track data and the National Hurricane Center (NHC) Atlantic data from 1979-2010. A TC is considered a candidate for ET if the track

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Although the raw resolution of the CFS reanalysis is 0.5°, at this resolution, the mass fields are noisy near TC, likely due to the method used to relocate the TC from its location in the 6 h forecast to the observed position in the analysis (e.g., Liu et al. 2000); therefore, the lower resolution 2.5° dataset, which shows no evidence of this issue, is employed. Given that this study mainly focuses on synoptic-to-planetary scale features, this choice of resolution should not be a limitation.
underwent recurvature at some point in its life, meaning it had an easterly component of motion, and the TC moved poleward of 20°N; the latter condition removed any TCs that drifted within the deep tropics, but never moved into the midlatitudes. For each of the remaining TCs, the reanalysis position is determined by finding the minimum in 1000 hPa geopotential height each 6 h; for a majority of times, the best track and reanalysis positions are within 1° of each other. At each time, the asymmetry parameter from the Hart (2003) cyclone phase space is calculated from reanalysis data. This parameter provides a measure of the thermal asymmetry across the cyclone and has been used to objectively identify the onset of transition. As in Hart (2003), the onset of transition is defined as when the asymmetry parameter exceeds 10 m and is hereafter referred to as “t = 0”. In each basin, the maximum number of ET cases within any 5° × 10° box is 34, which does not provide a robust composite to calculate wave packet properties due to the small number of cases. Instead, this study considers all cases where the onset of ET occurs between 30°-35°N and 120°-180°E in the WNP and 35°-40°N and 90°-30°W in the Atlantic; these latitude bands contain the largest number of ET onset within each basin. This choice results in 112 and 91 cases in the WNP and Atlantic, respectively. It is worth noting that this study considers all cases of ET within this location, regardless of whether the TC underwent baroclinic reintensification. Previous studies have suggested that the completion of ET is not necessary for a midlatitude response to occur (e.g., Harr and Dea 2009; Riemer and Jones 2010).

Figure 1 shows the number of ET and winter cyclones in each basin as a function of month. While the largest number of ET cases occur in September and October, the winter cases are mainly in February and March in the WNP and December-January in the Atlantic. The difference in timing between ET and winter cyclones results in different background states through which the wave packets develop and propagate; the implications of which are explored in greater detail in the next section.
a. Wave-packet analysis

The remainder of this paper will employ ensemble averaging of ET and baroclinic cyclone cases in each basin over a range of time lags. The relatively large number of baroclinic cyclone cases allow for straightforward averaging in an Earth-relative frame of reference; however, for the ET cases, all fields are shifted to a common longitude, which is defined as the longitude of the cyclone at the onset of ET. As pointed out in Hakim (2003), the ensemble averaging method has the advantage that it is relatively simple and does not integrate away time and amplitude information; however, the signal contained in the ensemble mean will degrade to zero for nonzero time lags due to different trajectories and velocities of individual events.

Properties of the wave packet at each time are analyzed using the methods outlined in Hakim (2003) and are summarized here. Wave packets are identified from the 300 hPa meridional wind field over a 250° longitude window centered on the maximum in the absolute value at a grid point. The peak of the packet is determined by fitting a polynomial to the six grid point extrema about the maximum value, while the local extrema and zero crossings are determined from the interpolated polynomial near the packet peak. The leading edge of the packet is determined by linear regression of the interpolated packet to an exponential profile, with the leading edge defined as 2.5 $e$-folding distance from the peak (8% of the peak value). Finally, the wavelength of the packet is determined by computing the distance between zero crossings and extrema.

3. Results

a. Western Pacific

Composites of the 300 hPa meridional wind during WNP ET and winter cyclone cases reveal important differences in the wave packet evolution prior to t=0. (Fig. 2). The dominant signal in the ET wave packet at $-48$ h is a $3 \text{ m s}^{-1}$ wind couplet centered near $155^\circ\text{E}$
with associated weak ridging in the PV field\(^2\) (Fig. 2a). By \(-24\) h, the amplitude of the meridional wind increases to 9 m s\(^{-1}\), with the corresponding undulation in the PV field suggestive of an amplifying ridge 5° to the east of the \(-48\) h position (Fig. 2c). At the onset of transition (0 h), the ridge amplitude increases further; however, the axis of this ridge has only moved 5° relative to the \(-24\) h position (Fig. 2e). Moreover, there is an indication of a downstream trough at 170°W, suggesting a nascent wave packet has developed. Overall, this result suggests that the process of ET produces an amplifying, but nearly stationary ridge, with little evidence of an upstream precursor disturbance. This result differs from Archambault et al. (2012) who showed that recurving cyclones are preceded by an upstream trough moving through the midlatitude flow. The difference between these two studies is likely due to how each study defines the lag time. Archambault et al. (2012) defined t=0 to be the time when the TC reaches its westernmost position, while this study defines t=0 to be the onset of transition; these two times can differ by 0-3 days depending on the case.

In contrast to ET cases, the winter cyclone composite is characterized by a predecessor wave packet prior to t=0. At t=\(-48\) h, there is a relatively weak (3 m s\(^{-1}\)) wave packet centered on Japan that subsequently amplifies as it moves eastward with time (Fig. 2b). Over the next 48 h, the meridional wind anomalies increase to 18 m s\(^{-1}\) during which time the packet moves 30° to the east (Fig. 2d,f). Similar to Hakim (2003), this pattern suggests that the winter cyclones tend to amplify a weak pre-existing wave packet that is moving through the midlatitude wave guide.

In addition to differences in how wave packets are generated during ET and winter cyclones, there are also subtle differences in how the packets evolve after t=0. For the ET cyclones, the wave packet is associated with an amplifying trough at 160°W at 24 h (Fig. 2g) and finally a nascent ridge with the axis over the west coast of the United States by 72 h (Fig. 2i,k) before vanishing over North America by 96 h (not shown). Although the

\(^2\)To facilitate comparison with the winter cyclone cases, the geography on the ET figures is oriented such that the cyclone position at t=0 matches the winter cyclone position at t=0
winter cyclone wave packet exhibits a similar eastward propagation, the southern edge of
the packet is characterized by refraction into the tropics starting at 24 h (Fig. 2h), which is
not present in the ET composites. Moreover, it appears that the winter cyclone wave packet
peak does not reach North America, while it does in the ET cyclones, which suggests that
on average ET cyclone wave packets propagate further east in the WNP. This difference is at
least partly due to the nature of the waveguide associated with these two types of systems.
The ensemble-mean meridional PV gradient in the ET cyclones is nearly constant across
the entire Pacific Ocean, which implies a zonally consistent waveguide. By contrast, the
winter cases are characterized by a higher meridional PV gradient to the west of the dateline
compared to the east. We hypothesize that the greater zonal variation in the meridional PV
gradient contributes to a greater range of group speeds for the winter cyclone cases, which
blurs the sample-average signal.

The ET and winter cyclone wave packet properties are objectively analyzed using the
methods outlined in section 2b to determine their wavelength, amplitude and group velocity.
Figure 3a,b shows the wave packets associated with ET and winter cyclones at t=0; these
figures are similar to what is obtained at t=−24 - +48 h (not shown). For both ET and winter
cases, the wave packets look similar to those observed in Hakim (2003). In particular, the
wave packet amplitude exhibits a good fit to an exponential profile with an abrupt westerly
edge, consistent with impulsive disturbances, which have an exponential zonal structure
(e.g., Swanson and Pierrehumbert 1994).

Comparing the packet peak amplitude confirms many of the aforementioned ideas of the
development and propagation of ET and winter cyclone wave packets. The ET packet peak
amplitude increases from 10 m s\(^{-1}\) at −24 h to 17 m s\(^{-1}\) 12 h after the onset of ET, then
decreases to 10 m s\(^{-1}\) by 48 h (Fig. 4a). By contrast, the winter packet amplitude is 3 m s\(^{-1}\)
higher than the ET wave packet at −24 h, peaks at 18.5 m s\(^{-1}\) at 0 h, and decreases at a
slower rate relative to the ET packet. While this result implies that winter cyclone wave
packets have greater amplitude than ET wave packets, this is potentially deceiving because
the climatological standard deviation in meridional wind is larger during the winter cyclones times (primarily February–March) compared to when ET occurs (September–October). To address this concern, the packet peak amplitude for each case is normalized by the climatological standard deviation in the 300 hPa meridional winds at the location of the packet peak for that day. Between $-24$ h and $+24$ h, the ET and winter cyclone normalized packet amplitudes are within 0.04 normalized units of one another; thereafter, ET and winter cyclone wave packets are of equal amplitude relative to climatology.

The group velocity calculations support the notion that the development of winter cyclones results in the enhancement of an existing wave packet, while ET leads to the generation of a new packet. Fig. 5 shows that the winter cyclone wave packet group velocity increases from 6 m s$^{-1}$ at $t=-24$ h to a maximum of 26 m s$^{-1}$ at 6 h, which is slower than the background flow of 40 m s$^{-1}$. The reduction in group velocity with time beyond 6 h likely reflects the decrease in the background flow across the basin, which could act to focus wave packets (e.g., Esler and Haynes 1999; Chang and Yu 1999; Hakim 2003). By contrast, the ET wave packet group velocity is either negative or zero until $t=0$, implying a nearly stationary wave packet during which time the amplitude is increasing (c.f., Fig. 4a). This result suggests that external forcing, such as latent heat release associated with the transitioning TC is critical and that the forcing has near zero or negative zonal velocity. Beyond 0 h, the group velocity increases to 15 m s$^{-1}$, except for the large spike at 24 h, which could be an artifact of the limited number of ET cases used in this compositing technique.

Figure 6 shows the packet peak wavelength for both the ET and winter cyclone cases. As suggested by Fig. 2, the ET wave packet has a wavelength that is 500–700 km longer than the winter cyclone cases at all lead times. These wavelength differences are likely due to the structure of the background flow through which the wave packets are traveling.

The different behavior of the ET and winter wave packets prior to 0 h suggests that different dynamical processes may be responsible for the generation and amplification of the wave packets. As stated earlier, both adiabatic and diabatic processes can contribute to
the amplification of the midlatitude flow. To investigate this possibility, ensemble-average lower tropospheric meridional temperature and moisture fluxes are computed at t=0 (other times are qualitatively similar) for both ET and winter cyclones. Fig. 7 shows the 900 hPa meridional heat flux, which is computed by multiplying the meridional wind deviation from climatology by the temperature deviation from climatology at each grid point. For both ET and winter cyclones, there is a maximum in meridional heat flux on the eastern side of the cyclone within the region of southerly geostrophic winds as implied by the geopotential height contours. Although the location of positive heat flux is similar in both sets of cases, the maximum in the winter cyclone heat flux is roughly twice the value of the ET cyclone. Moreover, the winter cyclones are also characterized by a greater spatial coverage of heat flux greater than 8 K m s$^{-1}$. Assuming that the largest temperature gradients are near the surface, this result suggests that, relative to ET cyclones, the winter cyclones are characterized by greater forcing for height rises and thus upper tropospheric ridge building via the quasi-geostrophic height tendency equation. Larger meridional heat fluxes might be expected for the winter cyclones given the enhanced climatological meridional temperature gradient during winter (not shown).

Whereas winter cyclones are characterized by larger meridional heat fluxes, the opposite is true for moisture fluxes at the same level (Fig. 8a,b). Here, moisture fluxes are computed in the same manner as heat fluxes, except that temperature perturbations are replaced by water-vapor mixing ratio deviations from climatology. Although the positive moisture flux area is similar in both cases, the maximum flux in the ET cases is 45 g kg$^{-1}$ m s$^{-1}$, compared to 27 g kg$^{-1}$ m s$^{-1}$ in the winter cyclones, suggesting that the ET cyclones are characterized by larger poleward moisture transport. Moreover, the ET cases are characterized by a maximum moisture flux convergence of $6.0 \times 10^{-5}$ g kg$^{-1}$ s$^{-1}$ on the poleward side of the moisture flux maximum, compared to $2.5 \times 10^{-5}$ g kg$^{-1}$ s$^{-1}$ in the winter cyclone. These differences imply that ET cases are characterized by greater latent heat release and forcing for height rises compared to winter cyclones. This result agrees with the idealized simulations
of Riemer and Jones (2010), who show that the initial downstream response to ET is mainly due to diabatic processes.

b. Atlantic

Composites of Atlantic basin ET and winter cyclone wave packets show many qualitative similarities to their western Pacific counterparts, with smaller ET amplitude. Figure 9 indicates that Atlantic ET is characterized by a nearly stationary amplifying ridge prior to the onset of ET, similar to WNP ET. In addition, there is a northerly wind signal that moves from western North America at −48 h to just upstream of the developing ridge at 0 h. By comparison, the winter cyclones are characterized by a weaker pre-existing wave packet located at 100°W at −48 h that subsequently moves east and amplifies with time (Fig. 9b,d,f).

Following the onset of ET, the wave packet moves eastward toward Europe; however, the statistically significant signal quickly decays, such that beyond 48 h, there is no signal in the 300 hPa meridional wind field. By contrast, the winter cyclone wave packet has a statistically significant signal that reaches Europe by 48 h and exhibits refraction into the tropics (Fig. 9h,j). Overall, these results suggest that, on average, ET wave packets have difficulty maintaining their amplitude as they move across the Atlantic Ocean relative to those associated with winter cyclones.

ET and winter cyclone wave packet properties are also computed within this basin. For winter cyclones, the peak packet amplitude is 21 m s$^{-1}$ at 0 h, then decays to half that value by 42 h when the packet reaches Europe, while the ET packet amplitude has a maximum value of 15 m s$^{-1}$ before decreasing to less than 10 m s$^{-1}$ by 36 h (Fig. 4b). Normalizing the amplitude by the climatological standard deviation indicates that the ET wave packet amplitude is about 10% less than winter cyclones.

Wave packet group velocities are consistent with the WNP results, which showed that winter cyclones enhance an existing wave packet, whereas ET packets are produced in situ.
The winter-cyclone wave packets have a group velocity that varies between 15–30 m s\(^{-1}\) throughout the period. By contrast, as was observed in the WNP cases, the ET wave packet group velocity is slightly negative until 0 h, which is likely related to the combination of the aforementioned midlatitude disturbance getting closer to the ridge associated with ET and the slow zonal motion of the forcing. Beyond that time, the ET packet peak group velocity is quite similar to the winter cyclones, suggesting its propagation properties are similar once the packet is of sufficient amplitude and detached from the forcing. Finally, the ET packet wavelength is greater than the winter cyclones for most times prior to 0 h, though they are quite similar thereafter (Fig. 6b).

In the lower troposphere, Atlantic basin winter cyclones have similar meridional heat fluxes relative to their western Pacific counterparts, while the Atlantic ET cyclones are characterized by smaller values compared to WNP ET. Although the spatial coverage of positive heat fluxes in the WNP and Atlantic ET composites are fairly similar, the maximum value in the Atlantic basin (20 K m s\(^{-1}\)) is 25% smaller than the western Pacific (cf., Fig. 7a,c). By contrast, the winter cyclones in both basins have similar areas of positive heat flux and maximum values (Fig. 7b,d); therefore, it appears that while the forcing for height rises in winter cyclones are similar in the two basins, the same is not true for ET. The lower heat fluxes for Atlantic ET would be expected to produce less forcing for height rises in the downstream ridge relative to WNP ET.

In addition to having weaker meridional heat fluxes, Atlantic ET cases also appears to have weaker moisture fluxes relative to the WNP cases (Fig. 8c-d). The maximum moisture flux in the Atlantic ET cases is 31 g kg m s\(^{-1}\), compared to 29 g kg m s\(^{-1}\) for Atlantic winter cyclones and 45 g kg m s\(^{-1}\) for WNP ET. This result suggests that, on average, Atlantic ET cases have smaller forcing for height rises due to diabatic heating, which could explain why the amplitude of ET wave packets in the Atlantic is smaller than winter cyclones.

Another reason for the lower amplitude in Atlantic ET wave packets vs. winter cyclones could be due to the case-selection criteria. Recall that the Atlantic composite includes all
TCs that underwent ET between 35°N and 40°N. For some of these cases, this happened while the TC was over land. By contrast, none of the WNP ET cases made landfall over an appreciable land mass, except for Japan. To determine whether these over-land ET cases bias the ensemble-mean ET wave packet, the composite calculations are repeated for all ET cases that remained over water prior to and during ET. The ensemble-mean wave packet for the over-water Atlantic ET cases is nearly identical to the ensemble-mean wave packet for all (not shown); therefore, land does not appear to be a factor in Atlantic ET having weaker wave packets.

Another possible factor for the difference in ET and winter cyclone wave packets is that ET is most frequent in August, September and October (ASO) when the background state could be less favorable for high-amplitude, long-lasting wave packets. The role of the seasonal cycle is tested by computing wave packet statistics for non-TC Atlantic cyclones that meet the cyclone criteria described in section 2 between 35°–40°N and 90°–30°W during ASO using the longitude-shifting procedure employed for ET cyclones. Wave packets associated with ASO cyclones have a group velocity of 10–12 m s\(^{-1}\), a wavelength of 3000 km, and packet peak amplitude of 24 m s\(^{-1}\) (normalized amplitude of 1.4 times climatology; not shown). These properties bear greater resemblance to winter cyclones relative to ET; therefore, it appears that the difference in Atlantic ET and winter cyclone wave packets cannot be attributed to the seasonal cycle alone.

4. Summary and Conclusions

This study poses the hypothesis that wave packets associated with the extratropical transition of tropical cyclones in the western Pacific and Atlantic basins are on average quantitatively indistinguishable from those associated with winter cyclones. This hypothesis is tested by computing ensemble averages of ET and winter cyclone cases from reanalysis data over a 32 yr period. The properties of the wave packet are then analyzed using the
techniques described in Hakim (2003).

For both the western Pacific and Atlantic basins, the wave packets associated with ET and winter cyclones exhibit many similarities, with the most significant differences apparent in the genesis and decay of the packet. Prior to the onset of ET, there is little evidence of an upstream wave packet in either the WNP or Atlantic basins. Instead, the development of the downstream ridge associated with ET appears to produce a wave packet in situ, which then propagates eastward once it escapes tropical forcing. By contrast, winter cyclone genesis appears to amplify a pre-existing wave packet that can be tracked backward in time through the midlatitude storm track. As a consequence, it is possible to reject the null hypothesis that there is no meaningful difference in the genesis of ET and winter wave packets.

Following the onset of ET and winter cyclone maturity, the wave packets exhibit similar characteristics of downstream propagation and eventual dissipation. In both basins, the group velocity and, to a lesser extent, the wavelength, are quantitatively similar, suggesting that once a packet matures, the behavior is similar in both cases. While WNP ET and winter cyclone wave packets have similar amplitudes relative to climatology, the amplitude of Atlantic winter cyclone packets is greater than ET. In addition, while WNP ET packets have a statistically significant signal that propagates further downstream relative to winter cyclones, the opposite is true in the Atlantic basin. These results suggest that wave packets associated with WNP ET and winter cyclones have similar structure and propagation characteristics, thus it is difficult to reject the null hypothesis that there is any meaningful difference. By contrast, Atlantic winter cyclone wave packets are strong and longer-lived than their ET counterpart, thus there are meaningful differences in this basin.

Some of the differences between ET and winter cyclone wave packets are likely related to the dynamics of packet development. In both the Atlantic and western Pacific basins, winter cases are characterized by larger tropospheric meridional temperature fluxes relative to ET cyclones, implying that winter cyclones have greater adiabatic forcing for geopotential height rises and thus wave packet amplification. On the other hand, WNP ET cases have larger
lower tropospheric meridional moisture fluxes and flux convergence relative to the winter cases. Assuming that moisture flux convergence correlates with latent heat release implies that diabetic processes are more important in wave packet amplification in ET relative to winter cyclones. In the Atlantic basin, the moisture fluxes during ET are comparable to winter cyclones and smaller than WNP ET, suggesting relatively smaller forcing for height rises and wave packet amplification, which could explain why Atlantic-basin ET wave packets are on average weaker than winter cyclones.

Although this work indicates that ET and winter cyclone wave packets have many similar characteristics, the differences in how these packets are produced could have important consequences on downstream predictability during each event. Given the relative dearth of in situ moisture observations over the lower-tropospheric ocean, models may have relatively large moisture analysis errors, which could translate into different latent heating rates and details in the downstream ridge amplitude during ET cases (e.g., Torn 2010). Moreover, since the packet is produced by the ET cyclone itself, TC track errors could also introduce uncertainty in the timing and amplitude of the subsequent wave packet (e.g., Harr et al. 2008; Anwender et al. 2008; Riemer and Jones 2010). By comparison, the winter cyclones appear to amplify pre-existing wave packets; therefore, provided the upstream disturbance is resolved by the current observation network, downstream predictability might not be as sensitive to details of winter cyclogenesis. Future work will evaluate the aforementioned hypothesis by comparing forecasts during ET events with those of winter cyclones.

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As in Fig. 7, but for the meridional flux of 900 hPa water vapor mixing ratio (g kg$^{-1}$ m s$^{-1}$).

As in Fig. 2, but for the Atlantic Basin.
Fig. 1. Number of winter baroclinic cyclone (black) and extratropical transition cases (gray) in the (a) western Pacific and (b) Atlantic basin used in this study as a function of month.
Fig. 2. Ensemble-mean time evolution of the Western Pacific extratropical transition (left column) and winter cyclone (right column) wave packets at 300 hPa. The heavy lines denote the meridional wind every 3 m s\(^{-1}\), with dashed indicating negative values, with the zero contour removed. In the left (right) columns, the thin lines denote the 330 K potential vorticity between 3-5 (1-3) PVU [1 PVU = 10\(^{-6}\) m\(^2\) K (kg s\(^{-1}\))\(^{-1}\)] each 1 PVU. The underlying map in the left column is oriented such that the composite center longitude at t=0 matches the winter cyclone box longitude.
Fig. 3. Western Pacific (a) ET and (b) winter cyclone wave-packet analysis at t=0. Solid lines denote the anomaly meridional wind (m s$^{-1}$) as a function of longitude, while the dashed lines are a linear fit to an exponential profile from the packet peak to 2.5 e-folding distances from the peak. (c) and (d) as in (a) and (b) but for Atlantic wave packets.
Fig. 4. Amplitude (m s\(^{-1}\)) of the ET (solid) and winter cyclone (dashed) packet peak (m s\(^{-1}\)) as a function of lag (h) for the (a) western Pacific and (b) Atlantic Basins. The gray lines indicate the amplitude normalized by the climatological standard deviation in meridional wind at that location and date.
Fig. 5. Zonal group speed (m s\(^{-1}\)) as a function of lag (h) for ET (solid) and winter cyclone (dashed) wave-packet peak in the (a) western Pacific and (b) Atlantic basins.
Fig. 6. Wavelength (km) of the ET (solid) and winter cyclone (dashed) wave packet as a function of lag (h) in the (a) western Pacific and (b) Atlantic basins.
Fig. 7. Ensemble-mean meridional temperature flux at $t=0$ for the ET (left column) and winter cyclone (right column) at 900 hPa (shading, K m s$^{-1}$). The heavy lines denote the ensemble-mean 900 hPa geopotential height every 20 m, with negative values dashed. The underlying map in the left column is oriented such that the composite center longitude at $t=0$ matches the winter cyclone longitude at $t=0$. 
Fig. 8. As in Fig. 7, but for the meridional flux of 900 hPa water vapor mixing ratio (g kg$^{-1}$ m s$^{-1}$).
Fig. 9. As in Fig. 2, but for the Atlantic Basin.