Determination of a Consistent Time for the Extratropical Transition of Tropical Cyclones. Part II: Potential Vorticity Metrics

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ABSTRACT

As a tropical cyclone moves poleward and interacts with the midlatitude circulation, the question of whether it will undergo extratropical transition (ET) and, if it does, whether it will reintensify or dissipate, is a complex problem. Uncertainties include the tropical cyclone, the midlatitude circulation, the subtropical anticyclone, and the nonlinear interactions among these systems. A large part of the uncertainty is due to a lack of an understanding of when extratropical transition begins and how it progresses. In this study, absolute potential vorticity and isentropic, or Ertel's, potential vorticity is examined for its ability to more consistently determine significant times (i.e., beginning or end) of the ET life cycle using the Navy Operational Global Assimilation and Prediction System gridded analyses.

It is found that isentropic potential vorticity on the 330-K potential temperature isentropic level is a good discriminator for examining the extratropical transition of tropical cyclones. At this level, a consistent “ET time” is defined as when the TC-centered circular average of isentropic potential vorticity reaches a minimum value. All 82 tropical cyclones moving into the midlatitudes meet this criterion. The completion of extratropical transition for the reintensifying cases is defined as when the storm exceeds an isentropic potential vorticity threshold value of 1.6 PVU at the 330-K potential temperature isentropic level. The success rate of this threshold value for the completion of extratropical transition for the reintensification cases is found to be 94.3% with a 27.6% false-alarm rate.

1. Introduction

The extratropical transition (ET) of tropical cyclones (TCs) can be described as the propagation of a tropical cyclone into the midlatitudes where it interacts with a preexisting midlatitude cyclone or a midlevel trough and transitions into an extratropical cyclone. As the TC moves into the midlatitudes, it encounters an increased Coriolis force, lower sea surface temperatures, and a dramatic increase in the vertical wind shear (Jones et al. 2003). These factors can act to decrease the maximum wind speed but increase the radius of gale-force winds (decrease in intensity but increase in strength), develop frontal-like precipitation, and increase wave heights.

The primary concern with ET events is flooding and wind damage, but they have also been known to cause bush fires in Australia on the dry side of the storm where strong winds exist and the increased wave heights can become dangerous for marine and shipping interests (Jones et al. 2003).

Although there has been a recent increase in the study of ET events globally, there is still a need to find a more accurate description of the TC and midlatitude trough interaction and to more consistently define the time at which the TC is no longer tropical, also called the “ET time.” Kofron et al. (2010, hereafter Part I) examines several methods that already exist in the literature for determining the ET time. These include the time when the TC becomes an open wave in the 500-hPa geopotential heights (Demirci et al. 2007), scalar frontogenesis (Harr and Elsberry 2000), and cyclone phase space (Evans and Hart 2003; Hart 2003; Hart et al. 2006). Part I found that while it may be possible to define an “ET completion”
time with each method and, with less accuracy or discrimination, an “ET onset” time, no method is consistent enough to be reliable using the Navy Operational Global Assimilation and Prediction System (NOGAPS) analyses, although the reliability of other analyses has not been determined. Of the three methods examined, cyclone phase space is perhaps the most widely used method for determining the ET onset and completion. However, Part I found that the phase-space definition is more useful for defining the completion of ET than is the onset of ET. This makes it a useful tool for retroanalysis of cases, but not so useful for real-time forecasting applications. In addition, it sometimes falsely classifies TCs as completing ET in the NOGAPS analyses and also does not distinguish between ET cases that reintensify or dissipate post-transition. Phase space requires the use of three parameters, which, when combined, provide a physically consistent view of the structure of any type of cyclone. However, no single parameter from the cyclone phase space can be used to define ET progression, and sometimes the progression of one parameter is not physically consistent with the other two.

The other definition examined in Part I, which appeared to have some utility for ET definition, was the “time of open wave” definition (Demirci et al. 2007). This definition met with similar successes and failures in distinguishing the different ET and non-ET classes as the cyclone phase-space parameters. While the open-wave definition has the advantage over cyclone phase space of being only one parameter to assess, it has the disadvantage of not being an automated procedure (i.e., there is no absolute value to calculate to determine ET onset). Thus, overall, Part I concluded that there is still a need to define ET time more consistently while better separating non-ET cases from ET cases and discriminating post-transition reintensification and dissipation.

A potentially useful tool for determining the ET time is potential vorticity (PV), which has been used to describe the evolution of TCs undergoing ET (Bosart and Lackmann 1995; McTaggart-Cowan et al. 2001; Thorncroft and Jones 2000). Similar to phase space, PV can be used to explain the structure of the TC and the balanced atmospheric flow surrounding the TC. This is possible using the invertibility principle in the absence of diabatic and frictional processes along a constant potential temperature surface. The principle assumes that there is geostrophic balance (Hoskins et al. 1985), which is not necessarily a good approximation for TCs, but may still be useful for transitioning TCs that are entering midlatitude regions in quasigeostrophic balance. Because TCs do in fact have intense diabatic processes confined to the core region and rainbands, their general structure should have higher PV values at lower levels with high PV gradients confined to the inner core (Jones et al. 2003; Shapiro and Franklin 1995). The values of PV decrease rapidly radially out from the center, and the PV gradients relax as well. The preexisting trough that the TC interacts with has higher PV values in the upper levels associated with the upper-level jet. Some more mature midlatitude cyclones may also exhibit higher low-level PV values along a low-level frontal boundary. Thus, the interaction of the TC and the trough may be described by the evolving midlevel isentropic PV.

In this paper, we will examine the usefulness of diagnostics based on potential vorticity to propose a new method for defining ET that can be easily understood by the operational and research communities as well as sufficiently explain the dynamics of ET. The hypothesis for this study is that the circularly averaged PV and isentropic PV may show a signal that separates the ET scenarios discussed in Part I. In section 2, the data and methods will be described, section 3 provides an analysis of the results, and a summary and conclusions are presented in section 4.

2. Data and methods

The 500-hPa geopotential height open-wave-centered data used for this study are the same as those used in Part I. Similar to Part I, the data are converted to an 83° by 73° storm-centered grid for all times. These grids are then used to calculate the potential temperature and isobaric, or absolute, potential vorticity, given by

\[
PV = \frac{1}{\rho} (\zeta + f) \nabla \theta, \tag{1}
\]

in a TC-centered framework. Here, the addition of the relative vorticity \( \zeta \) along a constant pressure level and Coriolis parameter \( f \) gives the absolute vorticity and \( \nabla \theta \) is the three-dimensional gradient of potential temperature, which is dominated in the vertical direction so that \( \nabla \theta \sim \partial \theta / \partial z \). The atmospheric density \( \rho \) can be calculated by substituting the ideal gas law:

\[
\frac{1}{\rho} = \frac{R_g T}{p}, \tag{2}
\]

where \( R_g \) is the gas constant of dry air, \( T \) is the temperature along a constant pressure level, and \( p \) is the pressure level. Using the assumption that \( \nabla \theta \sim \partial \theta / \partial z \), the hydrostatic approximation can be substituted into (1) to give Ertel’s isentropic potential vorticity (IPV):

\[
IPV = -g (\zeta + f) \frac{\partial \theta}{\partial p}, \tag{3}
\]
where $g$ is the gravitational constant and $\zeta_o$ is relative vorticity along a constant potential temperature surface. The PV and IPV are averaged within a circle of radius 500 km centered on the TC position. All PV and IPV values are expressed in potential vorticity units (PVUs), where $1 \text{ PVU} = 10^{-6} \text{ K m}^2\text{ kg}^{-1}\text{ s}^{-1}$. The reason for calculating both the isobaric PV and IPV will be given in section 3a in which an extension of the hypothesis for this study will be discussed.

The 82 TCs between 2003 and 2006 in the dataset are equivalent to those in Part I using NOGAPS reanalysis, which includes a bogus vortex at the TC center also discussed in Part I. The TCs are divided into four scenarios in which the TC undergoes ET with a trough to the northwest (42 cases) or northeast (16 cases) and re-intensifies (Harr and Elsberry 2000), undergoes ET and dissipates (8 cases), or recurs but does not undergo ET (16 cases). The TCs that undergo ET in conjunction with an upstream trough, or trough to the northwest, are further separated into cold-core transition cases (33 cases) and warm-seclusion cases (9 cases) determined by phase-space evolution (Hart et al. 2006). Average or composite fields are calculated for each scenario and examined for differences in storm evolution.

3. Results

a. Evolution of ET scenarios in the PV and IPV fields

During ET, the atmospheric flow around TCs changes dramatically and therefore PV, which is useful for describing balanced atmospheric flow, may be used to explain how and why the change occurs. The average potential temperature and PV for all 82 cases plotted versus pressure are shown in Fig. 1a. The potential temperature is averaged within a TC-centered 5° latitude by 5° latitude box, the PV is circularly averaged within a 500-km radius of the TC center, and the time series is centered on the 500-hPa geopotential height open-wave 0 h, or ET, time described in Part I. Figure 1a shows that, on average, the potential temperature decreases with time for recurring TCs and the largest PV values are within the 700- and 400-mb pressure levels. Below 700 mb, the potential temperature and PV values remain relatively constant prior to becoming an open wave. After the 0-h time, the average PV values reach a maximum between +12 and +48 h and the potential temperature decreases due to the general northward propagation of the TC. At the 850-mb level, the potential temperature decreases from 305 to about 295 K, and a decrease of about 10 K can be seen at all pressure levels up to approximately 250 mb. At about 200 mb the potential temperature remains mostly constant and above 200 mb the potential temperature increases with time, so the dynamic tropopause appears to be at or above 200 mb. Between 200 and 400 mb, the PV is constant until the 0-h time and then increases through the end of the period. The increase in PV in the upper troposphere is due to the interaction of the TC with the midlatitude feature.

On average, the 300-K isentrope does not exist above the 925-hPa pressure level and the tropopause is near 350 K (Fig. 1a); thus, IPV is plotted for the isentropic range of 305–350 K (Fig. 1b). The IPV values can be seen in Fig. 1a by tracing the PV along the potential temperature contours. However, the values in Fig. 1b differ slightly from those in Fig. 1a because of the conservative nature of Ertel’s IPV and are more appropriate to examine for an objective measure during ET. For this study, the troposphere can be divided into (Fig. 1a) the lower levels (below 700 mb), below 315 K before ET and below about 305 K at the end of the time series; the midlevels (between 700 and 450 mb), between 315 and 330 K before ET and between about 305 and 320 K post-transition; and the upper levels (450 mb to the tropopause), above 330 K before ET and above 320-K post-transition. Therefore, for the purposes of this study the low levels are defined as less than 315 K, the upper levels as above 330 K, and the midlevel of the atmosphere as the layer in between 315 and 330 K.

Figure 1b shows the mean and standard deviation of area-averaged IPV plotted on isentropes versus time where the 0-h time refers to the 500-hPa geopotential height open-wave time. In the lower levels (below 315 K), the average IPV is relatively low prior to 0-h time, especially below 310 K. At, and subsequent to, the 0-h time the IPV values in the lower levels increase and reach a maximum between +24 and +36 h. In the midlevels (between 315 and 330 K), the IPV values are relatively constant before the open-wave 0-h time (Fig. 1b), but there is an apparent minimum at the 330- and 335-K isentropic levels. After the open-wave 0-h time, the midlevel IPV begins to increase dramatically as the interaction with the midlatitude trough strengthens, and the increase occurs earlier at higher isentropic levels. The IPV in the upper levels (above 330 K) is slightly weaker than in the midlevels and remains below 1 PVU before the open wave 0-h time at 340 and 345 K. At the 350-K isentrope, the IPV values begin to dramatically increase at the open-wave time and are the earliest to increase of all the levels examined.

Figure 1c shows the average time series of grid-averaged sea level pressure minus the minimum central pressure for all 82 TC cases as well as for each of the separated 5 ET and non-ET scenarios described in Part I. The time series for all cases shows correspondence to the
FIG. 1. Time series centered on the 500-hPa geopotential height open-wave time for (a) the average of circularly averaged PV (PVU, shaded) and potential temperature (K, contour) for all recurving cases, (b) average (PVU, shaded) and standard deviation (PVU, contour) of the circularly averaged IPV for all recurving TCs, and (c) average difference of the grid-averaged sea level pressure and the minimum central pressure for each scenario and all TCs.
plot of absolute PV with little change in the relative pressure at the surface through the 0-h time, an overall increase to a maximum between +24 and +36 h when the near-surface PV maximizes, and a gradual decrease subsequently.

Examination of the composite plots of absolute PV and IPV for each of the ET and non-ET cases shows that there is potential discrimination among the different scenarios using these metrics. Figures 2–6 show composite plots of TCs that undergo the following: 1) ET with an upstream trough and a cold-core reintensification, 2) ET with an upstream trough and a warm-seclusion reintensification, 3) ET with a downstream trough and reintensification, 4) post-transition dissipation, and 5) recurvature but no ET. Figures 2a, 3a, 4a, 5a, and 6a show the pressure versus time plot of potential temperature (dashed contour), the isobaric PV (solid contour), and the average wind magnitude within a TC-centered 500-km radius (shading). Figures 2b, 3b, 4b, 5b, and 6b show the u and v 850–200-mb wind shear components. The potential temperature versus the IPV anomaly (the total IPV field minus the −72-h cold-core upstream trough IPV) is plotted in Figs. 2c, 3c, 4c, 5c, and 6c.

1) SCENARIO 1: COLD-CORE REINTENSIFICATION WITH AN UPSTREAM TROUGH

For the TCs that interact with an upstream trough and undergo cold-core reintensification the PV decreases through the mid and upper levels of the atmosphere between −24 and +36 h and subsequently increases (Fig. 2a). Both components of wind shear increase steadily after −24 h as the midlatitude trough begins to encroach on the TC, reach a maximum about 36 h after the 0-h time, and subsequently decrease (Fig. 2b). The increase in shear and the decrease in absolute PV values correspond to the effect of the encroaching midlatitude trough on the TC and its immediate environment (within 500 km). The average central pressure of the ET system reaches a minimum at +48 h (Fig. 1c) during the period that the PV is increasing through much of the atmospheric column (Fig. 2a). This is when the vertical wind shear begins to decrease (Fig. 2b), suggesting that the ET cyclone reaches a maximum intensity as it begins to occlude. Below 700 mb, the average PV values remain nearly constant with a slight rise at later times. This consistency and the lower wind values suggest that the PV at the very lowest levels is not directly impacted by the midlatitude trough.

Before the open-wave time, the IPV anomalies are near 0 with a small increase at the 350-K level at +48 h (Fig. 2c). Between 330 and 345 K, the IPV anomalies are slightly negative around the 0-h time because of the outflow from the TC, but then increase as the midlatitude trough interacts with the TC. Below 325 K, the IPV anomalies increase slightly throughout the time series until the trough begins to interact with the TC in the lower levels. The maximum low-level IPV anomalies occur near +48 to +60 h corresponding with the minimum central pressure in Fig. 1c.

2) SCENARIO 2: WARM-SECLUSION REINTENSIFICATION WITH AN UPSTREAM TROUGH

Figure 3 shows the same composite panels for the upstream trough ET and warm-seclusion reintensification. The early evolution of absolute PV and wind (Fig. 3a) is quite similar to that in Fig. 2a where the PV has a relative maximum in the midlevels, low values in the upper levels below the dynamic tropopause, and steady values in the lower levels. The wind is also relatively weak through the atmosphere until the 0-h time when the wind shear also increases (Fig. 3b). However, after the 0-h time, the evolution differs considerably from that in Figs. 2a,b. The upper-level trough is stronger and the trough extends lower in the atmosphere. The mid- and lower-level PV values increase rapidly for the first 24 h after the 0-h time until a tall vertical core of high PV values (>1.80 PVU) extends throughout the troposphere from the upper-level trough to the surface (Fig. 3a). The increase in the PV values is consistent with not only the rapid intensification of the cyclone throughout the atmosphere, but also the strength that the surface cyclone generally achieves during warm-seclusion ET (e.g., Fig. 1c) compared with other types of ET. The magnitude of the wind and the wind shear in the upper levels increase earlier and more rapidly, but to a lesser extent than the wind in Fig. 2a, and the values drop off again very rapidly so that by +36 h they are insignificant through a deep layer compared with Fig. 2a. Similar to the cold-core ET cases, the maximum intensity is achieved after the wind shear decreases, which is between +24 and +36 h (Fig. 1c).

However, the entire development is indicative of thewarm-seclusion ET and demonstrates a very different evolution to that of the cold-core ET evolution (Figs. 2a and 3a). The IPV anomalies relative to the −72-h cold-core case are plotted in Fig. 3c and show that compared to the cold-core upstream trough cases the average central pressure of the warm-seclusion TC prior to ET is lower, the reintensification occurs more rapidly, and the final surface intensity of the extratropical cyclone after ET is greater.

3) SCENARIO 3: REINTENSIFICATION WITH A DOWNSTREAM TROUGH

Figure 4 shows the same plots for the TC that interacts with a downstream trough and reintensifies. The early evolution of absolute PV and wind (Fig. 4a) are also very
FIG. 2. Time series centered on the open-wave time for upstream cold-core trough reintensification ET cases: (a) average magnitude of wind (shaded, m s$^{-1}$), average potential temperature (dotted contour, K) within a 2° latitude × 2° longitude box around the TC center, and circularly averaged absolute PV (solid contours, PVU); (b) 850–200-mb wind shear for the $u$ (solid, open circle) and $v$ (dashed, closed circle) components of wind; and (c) IPV anomaly (PVU).
FIG. 3. As in Fig. 2, but for upstream trough warm-seclusion reintensification cases.
FIG. 4. As in Fig. 2, but for downstream trough reintensification cases.
FIG. 5. As in Fig. 2, but for post-transition dissipation cases.
FIG. 6. As in Fig. 2, but for TCs that recurve into the midlatitudes but do not undergo ET.
similar to that in Fig. 2a where the PV has a relative maximum in the midlevels, low values in the upper levels below the dynamic tropopause, and steady values in the lower levels. As the open-wave time approaches, the wind increases considerably compared to the upstream trough ET cases (Fig. 4a) and the wind shear also increases (Fig. 4b). After the open-wave time, the evolution of the downstream trough ET cases differs from the cold-core upstream trough cases. The $u$ component of the wind shear (Fig. 4b) is much greater than the $v$ component because the TCs are advected by strong zonal flow around the base of the downstream trough (Harr and Elsberry 2000; Part I) and the total wind shear increases to over 35 m s$^{-1}$ by +48 h before beginning to weaken. The evolution of the PV (Fig. 4a) is more similar to the cold-core reintensification ET cases (Fig. 2a). However, the time that the interaction between the TC and upper-level trough begins is about 12–24 h later because of the delay that occurs as the TC advects around the base of the downstream trough and the reintensification, although delayed compared with the upstream cold-core cases, appears to happen much more rapidly and produce a stronger average ET cyclone (Figs. 1c and 4a). In fact, the ET cyclone has not reached the final maximum intensity by the end of the calculated time series. The magnitude of the advecting flow can be seen in Fig. 4a to be much stronger than that of both the northwest trough cold-core and warm-seclusion reintensification cases (Figs. 2a and 3a). The corresponding IPV anomalies relative to the −72-h cold-core case are plotted in Fig. 4c and show that compared to the cold-core upstream trough cases the average downstream trough ET TC prior to 0 h is weaker, the reintensification occurs later because of the delay that occurs as the TC advects around the base of the trough before beginning its interaction with the midlatitude feature, and the final surface intensity of the extratropical cyclone after ET is higher (Fig. 4c).

4) SCENARIO 4: DISSIPATING ET CASES

The evolution for post-transition dissipation ET cases is shown in Fig. 5. The structure of the absolute PV in Fig. 5a for the post-transition dissipation cases looks quite similar to the structure of the reintensification ET cases with maximum values in the midlevels and low values in the upper levels as the TC approaches the midlatitude zone. However, the approaching upper-level trough is much shallower than for the reintensification cases, indicating that the midlatitude trough is located farther north relative to the TC for these cases (Fig. 5a). In addition, the wind shear, particularly the zonal component (Fig. 5b), continues to increase with time throughout the period unlike the reintensification cases. This implies that the TC remains under the influence of the increasing wind shear associated with the midlatitude westerlies and never interacts with the upper-level trough, which is located farther to the north, and so dissipates (Fig. 1c). The IPV anomaly plot in Fig. 5c also suggests that the upper-level trough is located farther to the north relative to the TC than for the reintensification ET cases. In general, the posttransition dissipation cases are located in the strong shear zone, but not in the favorable upper-level jet entrance region for reintensification (Figs. 2c and 5c).

5) SCENARIO 5: RECURVING TCs THAT DO NOT UNDERGO ET

Finally, the evolution for recurring non-ET cases can be seen in Fig. 6. The absolute PV values for non-ET cases (Fig. 6a) are slightly weaker at early times compared with the ET cases (Figs. 2a, 3a, 4a, and 5a), with a small maximum near the 0-h time. However, a distinguishing characteristic is that the PV remains stacked through a deep layer even as it weakens slightly after the ET time. The upper-level trough appears slightly later than for the dissipating ET cases and is even weaker and shallower in the composite (Fig. 6a), indicating that the midlatitude trough is located quite far north relative to the TC. The wind magnitudes are weaker than those for the dissipating cases; they reach a maximum relatively early, between +24 and +36 h, and subsequently weaken (Fig. 6a). Similarly, the wind shear exhibits a higher zonal component than meridional component, but the overall values are lower than those for the dissipating cases (Figs. 5b and 6b). The lower wind magnitudes combined with lower vertical wind shear (Figs. 6a,b) imply that the recurving TCs remain farther south and do not move as far north into the baroclinic zone as those TCs that undergo ET and either reintensify or dissipate. After the 0-h time, some cases even move back to the south out of the midlatitude westerly zone and reintensify slightly (e.g., see Fig. 1d in Part I). The IPV anomaly field highlights the case that the upper-level trough is located farther to the north relative to the TC (Fig. 6c) than any of the ET cases. In general, the recurving non-ET cases recurve into the southernmost portion of the midlatitude shear zone and as they begin to recurve back out of the midlatitudes, the shear values decrease.

b. Defining the ET time

The time series of IPV at different isentropic levels separated into the different types of ET are shown in Fig. 7. One representative isentropic level is shown for the lower and upper levels (305 and 340 K, respectively) and plots every 5 K for the midlevels are shown (315–330 K).
There is a distinctive increase in the IPV values for the reintensification cases (Figs. 7a–c) after the open-wave 0-h time. However, there is no corresponding increase for the dissipation cases (Fig. 7d) or the recurring non-ET cases (Fig. 7e) in the lower and midlevels. Furthermore, the IPV decreases slightly to a minimum near the 0-h time before the increase begins for the post-transition reintensification scenarios. In contrast, the average IPV for the posttransition dissipation and recurring non-ET scenarios increase slightly before the open-wave 0-h time and decrease after the open-wave 0-h time in the lower and midlevels until at least +36 h. There is a slight increase earlier in the (upper level) 340-K IPV, but this is due to a weak interaction with the midlatitude trough in the upper levels.

The discussion of the IPV evolutions from Figs. 2–7 suggests it may be possible to use the midlevel IPV to determine the ET time and separate the reintensification cases from the dissipating or non-ET cases. The 330-K isentropic is examined for this purpose. From Figs. 2–6, it is found that the 330-K isentropic level begins on average at 500 mb and ends in the time series below the 300-mb level for the reintensification cases, around 450 mb for post-transition dissipation cases, and just above 400 mb for non-ET cases. The differing heights in the atmosphere for each of the categories are simply due to the
differing proximities of the TC relative to the trough. Therefore, TCs that remain farther south in the Northern Hemisphere will have isentropic surfaces lower in the atmosphere than TCs that are displaced farther north.

Figure 8a shows the plots for each scenario of the 330-K IPV centered on the original open wave 0-h time. The minima in Fig. 8a are representative of the time that the IPV reaches a minimum for each TC in the respective scenario. For example, the upstream trough ET cases reach a minimum on average very close to the open-wave time, whereas the downstream trough ET cases reach a minimum on average about 12 h later (Fig. 8). The dissipation and non-ET cases reach minima at later times on average because the storms continue to weaken without interacting with any midlatitude features.

The data for individual storms may sometimes contain two IPV minima near the ET time. Therefore, an additional criterion is set such that the IPV minimum must occur within 24 h of a first-guess open-wave time. This is consistent with the study of cyclone phase space (Hart 2003) in Part I where only a few TCs in the dataset begin ET earlier than 24 h before the open-wave 0-h time using the asymmetry parameter. After each IPV minimum is found, the time series are then recentered on the IPV minimum and the results are shown in Figs. 8b,c. In the new time series, the IPV minimum for each scenario is more consistently located at the 0-h time, and the subsequent increase in IPV is clearly shown. In addition, the IPV increases substantially more for the reintensification cases than for the dissipation and non-ET cases (Fig. 8c). Although there is a slight increase in the IPV after the 0-h time for the post-transition dissipation and non-ET cases in Figs. 8b,c, this is due to a very slight influence from the trough in the former scenario and to several storms reintensifying as tropical cyclones rather than completing ET in the latter scenario.

c. Discriminating the true ET cases from dissipaters and non-ET cases

All of the 82 cases examined reach an IPV minimum at the 330-K level, which currently defines the ET time. However, only 58 of the 82 cases reintensify post-transition while the other 24 either dissipate after moving into the midlatitudes or recurve back out of the midlatitudes into the tropics. Klein et al. (2000) suggest that the post-transition dissipation cases should not be considered to
complete ET and therefore should be added to the non-ET cases for the purposes of discriminating cases that complete ET. It would be nice to find a criterion that separates the true ET cases from the other cases (i.e., the dissipation cases and the non-ET cases). Reexamination of Fig. 7 indicates that there is somewhat different behavior captured at the different isentropic levels. First, prior to the open-wave 0-h time, all isentropic levels have very similar values for all scenarios, and all values vary only slightly. Subsequent to the open-wave time, the IPV values spread considerably for the individual scenarios, most notably for the reintensifying scenarios. For example, in almost all five scenarios (warm seclusion is the exception) the low-level (305–315 K) IPV values tend to vary only slightly through the time series compared with the upper-level (330–340 K) values. In addition, the 315-K IPV values tend to rise slightly for the intensifiers post-ET time and decrease for the dissipaters and non-ET scenarios. Figure 8d shows the recentered 315-K IPV (recentered on the 330-K IPV minimum), plotted relative to its 0-h value. Interestingly, while the reintensifying scenarios increase significantly post 0 h, the dissipating cases increase little and the non-ET cases decrease relative to the 0-h time. The results in Fig. 9a are intuitive. Because the 315- and 330-K IPV values for all scenarios are fairly similar prior to the ET time, the plots lie on top of each other and it is difficult to choose decision boundaries to separate any particular groupings. However, after the ET time (Fig. 9b), the behavior of the reintensifying groups compared with the other two is quite different. As the 330- and 315-K values increase from their 0-h time values, the paths of the reintensifying groups move rapidly to the upper right of Fig. 9b (i.e., increasing values of both the 315- and 330-K IPV), whereas the paths of the dissipating and non-ET groups move to the left of the panel (decreasing values of the 315-K IPV). Thus, while the ET time could be set by the minimum in the 330-K IPV values, the scenarios could be separated by the subsequent behavior of the 315-K IPV values (significantly increasing for the reintensifying scenarios and decreasing for the dissipating and non-ET scenarios).

Alternatively, Fig. 8b shows clearly that a threshold IPV value could be set at the 330-K level that separates the different scenarios at the completion of ET. The cases that exceed this threshold would be considered to have completed ET and those that do not exceed the threshold would have failed to complete ET. The success rate of the threshold could also be considered as the success rate of using potential vorticity as a metric for determining the life cycle of a TC undergoing ET. Based on Fig. 8 for the average recentered IPV values, a useful threshold could be set between 1.25 and 2.0 PVU at the 330-K level. Thus, three metrics (1.25, 1.5, and 2 PVU) will be examined to determine which threshold works best. The results for each metric can be seen in Table 1. For a threshold of 1.25 PVU at the 330-K isentropic level, the success rate for correctly capturing the reintensification cases is 91.4% while 25% of the post-transition dissipation and non-ET cases are captured. While these rates appear to successfully discriminate ET cases well, several of the IPV minima never drop below the threshold value. Therefore, the 1.25-PVU threshold may not be appropriate. A threshold value of 2 PVU successfully captures only 81.0% of the post-transition reintensification cases but successfully rejects all of the post-ET dissipation and non-ET cases. The middle threshold value of 1.5 PVU appears to work best by capturing 84.5% of the post-transition dissipation and non-ET cases. In addition, all of the individual IPV minimum values are below this threshold value of 1.5 PVU.
A sensitivity test for properly determining the ET completion time IPV threshold value is the receiver operator characteristic (ROC) curve, which tests the accuracy of each threshold value at each isentropic level. The ROC curves for IPV thresholds set between 0.5 and 2.0 PVU are shown in Fig. 10a for IPV calculated on the 320-, 325-, and 330-K isentropic levels. The true-positive rate is calculated as the number of reintensification cases that successfully complete ET (true-positive cases) divided by the total of true-positive cases and false-negative cases, which are the post-transition dissipation and non-ET cases that are incorrectly classified as ET completion cases. The false-positive rate is calculated as the number of reintensification ET cases that are found to not complete ET (false-positive cases) divided by the sum of the false-positive cases and the true-negative cases, which are post-ET dissipation and non-ET cases that successfully do not complete ET. These rates are calculated at every 0.1 PVU between the range of IPV values given above. In Fig. 10a, it can be seen that the curves are noisy, so the accuracy of each value is also plotted in Fig. 10b. The accuracy of the thresholds at 330 K is no less than 80% accurate using any IPV threshold and is generally higher than the 320- and 325-K levels, which shows that the 330-K isentropic level is statistically appropriate for determining the completion of ET. A threshold value of 1.6 PVU successfully captures 94.3% of the post-ET reintensification cases with a false-alarm rate of 27.6%. In addition, an overall accuracy of 86.6% is found at 1.6 PVU using the ROC curves, which is very close to the combined success rate of 85.4% for the 1.5-PVU threshold in Table 1. Therefore, the 1.6-PVU threshold at the 330-K isentropic level appears to be the best value for defining the ET completion time and discriminates between reintensification cases and dissipation and non-ET cases.

### Table 1. Table of success and fail rates (%) of the defined ET completion time using isentropic potential vorticity (PVU) at the 330-K potential temperature isentropic level. The threshold values are reached after the minimum time in the IPV time series.

<table>
<thead>
<tr>
<th></th>
<th>1.25 PVU</th>
<th>1.5 PVU</th>
<th>2.0 PVU</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Success</td>
<td>Fail</td>
<td>Success</td>
</tr>
<tr>
<td>NW trough cold core</td>
<td>30 3</td>
<td>28 5</td>
<td>27 6</td>
</tr>
<tr>
<td>NW trough warm seclusion</td>
<td>9 0</td>
<td>8 1</td>
<td>8 1</td>
</tr>
<tr>
<td>NE trough</td>
<td>14 2</td>
<td>13 3</td>
<td>12 4</td>
</tr>
<tr>
<td>Positive (58 cases)</td>
<td>91.4 8.6</td>
<td>84.5 15.5</td>
<td>81.0 19.0</td>
</tr>
<tr>
<td>Post-ET dissipation</td>
<td>1 6</td>
<td>1 6</td>
<td>0 8</td>
</tr>
<tr>
<td>Non-ET</td>
<td>5 11</td>
<td>2 14</td>
<td>0 16</td>
</tr>
<tr>
<td>Negative (24 cases)</td>
<td>25.0 75.0</td>
<td>12.5 87.5</td>
<td>0 100.0</td>
</tr>
</tbody>
</table>

**d. Recentered IPV**

Having determined that the 330-K IPV “minimum” time for each storm is an appropriate ET time, the storms are recentered on the 330-K IPV minimum. Figure 11 shows the averaged cross sections for each scenario using the 330-K minimum time as the 0-h time instead of the open-wave time for $\pm 48$ to $+48$ h. Each category in Fig. 11 looks similar to the corresponding scenario image discussed in section 3a, but it is more evident that the IPV for reintensification cases (Figs. 11a–c) weakens in the upper levels before the midlevels as the average TC approaches ET time. The low-level IPV actually increases throughout the time series, and the mid- and upper-level IPV rapidly increases after the ET time. For upstream trough warm-seclusion reintensification cases (Fig. 11b), the average IPV increases more rapidly in the lower levels than the upstream trough cold-core reintensification (Fig. 11a) and downstream trough ET (Fig. 11c) cases because the warm-seclusion storms have a stronger low-level vortex entering into the transition period and generally higher IPV values near the center.
through the transition period. Also, it is clear that the low and midlevel IPV increases slightly later for the downstream trough re-intensification ET cases compared with the upstream trough ET cases. For the post-transition dissipation and non-ET cases (Figs. 11d,e), the IPV decreases at all levels before the IPV ET time. The weakening occurs more strongly for the non-ET case average than for the post-ET dissipation case average and it appears that the average non-ET case is rather weak compared with all other scenarios. After the 330-K IPV minimum time, the non-ET IPV values increase rapidly in the upper levels similar to the other scenarios as the upper-level trough approaches. However, similar to Fig. 6, the upper-level trough is far north of the TC and the increasing IPV values are restricted to the upper atmosphere. Finally, there is evidence in Fig. 11e that some of the non-ET cases do re-intensify post-transition as early as 36 h after the ET time as a separate entity from the upper-level trough. This is probably because some of the non-ET cases re-intensify as TCs after recurving back into the tropics or into favorable areas for tropical development.
4. Conclusions

Several methods for determining the ET time were examined by Part I, but the results show that the success rate of capturing the reintensification cases was not consistent and too many of the post-transition dissipation cases and non-ET cases were improperly found to undergo ET when in reality they did not. This study attempts to alleviate some of the issues with those methods for determining the life cycle of a TC undergoing ET by examining PV and IPV as an option for numerically defining the ET onset time and ET completion time, which correspond to the Klein et al. (2000) transformation beginning and end times, respectively. It is found from the time series plots of the circular average of absolute PV that the levels at which the ET is most important are in the mid and upper levels of the atmosphere. These levels translate into isentropic levels as the range between 305 and 350 K. Further investigation of the time series plots of IPV reveal that the best level for examining ET is at the 330-K potential temperature level because differences in how the low-level atmosphere develops compared with how the upper-level atmosphere develops, including the interacting midlatitude trough, are well captured for all scenarios. In general, TCs that enter into the midlatitude environment and undergo ET show a dramatic increase in the 330-K IPV while TCs that do not interact with midlatitude features or TCs that dissipate post-transition do not encounter increased IPV values.

A new ET time is determined by using a local minimum value of 330-K IPV as the “IPV defined” ET onset time. A difficulty arises with this definition of the ET time because (i) there may be more than one minima in any individual recurring TC; and (ii) there is a reliance on the data at the time before and after the minimum, which means that the ET time cannot be determined until after just after it has occurred or must be determined using forecast data. All 82 recurring TCs begin ET using this definition. Examination of the 315-K IPV centered on the 330-K IPV minimum time shows differentiation between the scenarios that complete ET (reintensifying classes), and those that do not (dissipating and non-ET classes).

An additional way to discriminate these classes of recurring TCs is to numerically define an “ET completion” time. The ET completion time appears to be well defined by the 300-K IPV threshold value of 1.6 PVU. Because there is no need to rely on forecast data or later analyses, the completion of ET can be determined at the moment it occurs. However, it is noted that if the goal was to predict the completion of ET, then there would be a need to use forecast fields to determine the threshold. The success rate at which this threshold performs is higher than for the previous methods discussed in Part I, and it successfully discriminates between the post-transition reintensification cases and the post-transition dissipation cases and non-ET cases. The likely cause of reaching the threshold or not has to do with the strength of the interaction between the remaining TC and the midlatitude trough. The dissipation cases and non-ET cases do not interact as strongly with the trough and therefore do not reach the threshold value. The average increase in IPV values for these two scenarios is a result of a weak interaction or intensification as a tropical feature after reaching the IPV minimum. Finally, although not examined in this study, the reintensification stage completion could be in conjunction with the maximum value of IPV reached, which in some cases could actually be at the same time as the completion of the transformation stage.

Future work on whether IPV is a useful metric for determining a consistent ET time and providing discrimination among different types of ET will include increasing the number of years in the analysis so that a more robust training set of TCs can be used. In addition, the work will be extended using analyses that do not include a TC bogus. For this study, it is assumed that the TC bogus in the NOGAPS data has little effect on the data other than better initializing the TC center. Whether this is true may be determined by the use of nonbogus analyses. Furthermore, we would like to see if the values of IPV for discrimination in the Atlantic and western North Pacific also apply to the eastern North Pacific and Southern Hemisphere. Other future work will also include finding a consistent and meaningful ET time by use of satellite imagery, which will remove the current reliance on model analyses. The goal of all research is to eventually design a forecast model based on statistics that can successfully forecast the ET life cycle.

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REFERENCES


