Simulations of the Extratropical Transition of Tropical Cyclones: Contributions by the Midlatitude Upper-Level Trough to Reintensification

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ABSTRACT

The extratropical cyclone development processes during the reintensification stage of an extratropical transition from a tropical cyclone (TC) are described using numerical simulations. Three control simulations without a tropical cyclone present examine the extratropical cyclogenesis associated with upper-level troughs that are characterized as weak, moderate, and strong. When no tropical cyclone is included in the simulation, the minimum surface pressures attained with the weak, moderate, and strong troughs are 1003, 991, and 977 mb, respectively. In all three cases, the low tilts northwestward with height during intensification, and the rainfall pattern and eventual occlusion are representative of classic extratropical cyclone development.

The interactions of a tropical cyclone with each of the three midlatitude circulation patterns are compared with the control simulations to illustrate the contributions to the extratropical transition of the tropical cyclone. In the three trough-with-TC cases, the minimum surface pressures were almost identical (967, 965, and 959 mb). Thus, the final intensity of the extratropical cyclone is not only related to the strength of the upper-level trough but must also be related to the structure of the basic midlatitude environment. The proper phasing of the tropical cyclone with the midlatitude trough results in substantial enhancement of the upper-level divergence. In addition, higher $u$ values in the lower troposphere associated with the tropical cyclone remnants are absorbed in the developing extratropical cyclone. The lifting of this moist air results in precipitation that is greater in both amount and areal extent, which enhances extratropical development when compared with the control cases. Based on these simulations, an important conclusion is that a weak midlatitude trough interacting with tropical cyclone remnants may have as much potential to intensify, as does a moderate or strong trough, and may have longer periods of rapid intensification.

A development potential parameter based on the three main factors in the Petterssen development equation (upper-level divergence, midlevel positive vorticity advection, and low-level temperature advection) is calculated for all simulations. The strength and areal extent of the development parameter has utility in predicting where and whether extratropical cyclogenesis will occur during the reintensification stage of extratropical transition.

1. Introduction

The extratropical transition (ET) of tropical cyclones (TCs) has been described by Klein et al. (2000) as a continuous process that can be divided into two stages. The first stage, which they call transformation, includes the initial structural changes that a recurving tropical cyclone undergoes as it begins to interact with a pre-existing baroclinic zone and the associated vertical wind shear. Common characteristics of this interaction include lower-tropospheric thermal advection by the tropical cyclone circulation, ascent and descent along the tilted isentropic surfaces of a baroclinic zone, development of a vertical-motion dipole, dispersal of the upper-level tropical cyclone warm core, enhancement of the low-level tropical cyclone warm core (Ritchie and Elsberry 2001), and lower-tropospheric frontogenesis. Klein et al. (2000) consider the transformation stage finished when the transformed tropical cyclone is embedded in cold, descending air within the baroclinic zone. Klein et al. (2000) also defined a second stage of ET called reintensification, in which the transformed tropical cyclone either dissipates or else reintensiﬁes if the upper troposphere is favorable for extratropical cyclogenesis [i.e., upper-tropospheric divergence and positive vorticity advection (PVA) are superposed above low-level baroclinity as described by DiMego and Bosart (1982) and Sinclair (1993)].

Klein et al. (2000) noted many similarities in the tropical cyclone structure changes that occurred during the transformation stage in the 30 western North Pacific
cases they studied and concluded that transformation processes are a natural consequence of a tropical cyclone moving poleward and interacting with a baroclinic zone. However, the subsequent reintensification stage was found to be very dependent on the details of the mid-latitude circulation structure and how the transformed storm interacted with that midlatitude circulation. In particular, not all poleward-moving tropical cyclones undergo reintensification. For those that do, the ability of forecast models to accurately predict the rate of intensification and future track of the storm is at times poor, with large discrepancies from forecast to forecast even by the same model (P. Harr 2000, personal communication). This led to the Harr and Elsberry (2000) hypothesis that the details of the structure of the transformed tropical cyclone have relatively little influence on the intensification processes; rather, it is the structure of the midlatitude environment that determines the rate and nature of reintensification. If the interaction of the transformed tropical cyclone with the midlatitude environment is poorly handled by the forecast model, a poor reintensification forecast results. This circumstance was also noted by McTaggart-Cowan et al. (2001) in their case study of Hurricane Earl of 1998. By using piecewise potential vorticity inversion techniques, they were able to isolate the relative importance of the hurricane remnants and the upstream midlatitude upper-level trough in the subsequent reintensification of the extratropical cyclone. Although their technique was different from that of Klein et al. (2002), their conclusion that the hurricane remnants have little impact on the subsequent reintensification was similar.

Prior studies of ET (DiMego and Bosart 1982; Sinclair 1993; Foley and Hanstrum 1994; Harr and Elsberry 2000; Harr et al. 2000) have described reintensification of the transformed tropical cyclone in terms of Type-B development, following Petterssen and Smebye (1971). In this framework, low-level cyclone development occurs when an area of midlevel PVA becomes superposed over a low-level frontal region. When the midlevel PVA increases over the low-level frontal zone, low-level thermal advection can contribute significantly to the increase in low-level vorticity. When the thermal field is distorted through circular motion, as may occur during transformation of a tropical cyclone (Klein et al. 2000), the thermal contribution becomes significant (Petterssen 1956).

Klein et al. (2002) reason that if Petterssen–Smebye Type-B cyclogenesis is the process by which a transformed tropical cyclone reintensiﬁes during ET, the tropical cyclone (and low-level patterns of cold and warm advection) must translate poleward to phase with midlevel PVA that exists in a midlatitude circulation pattern. This led Klein et al. to define a “development region” based on their 30-case sample as the region where any one of the three main factors in the Petterssen development equation (upper-level divergence, midlevel PVA, and low-level temperature advection) attains threshold values set as $3 \times 10^{-5} \text{ s}^{-1}$, $20 \times 10^{-10} \text{ s}^{-2}$, and $\pm 10 \times 10^{-9} \text{ K s}^{-1}$, respectively. Klein et al. (2002) assigned a value of 1 for each parameter that reached its threshold value, so a maximum possible development value of 3 indicates that the region has strong extratropical development potential. Depending on the phasing of the transforming tropical cyclone with the development region enhanced by the upper-level trough, a simple predictor of the strength of reintensification of the transitioning cyclone may be possible.

The purpose of this study is to investigate the role of the varying strength of the midlatitude, upper-level trough on the reintensification stage of ET. McTaggart-Cowan et al. (2001) have documented the importance of the midlatitude trough in the reintensification stage of Hurricane Earl by selectively removing the trough from the initial conditions using potential vorticity inversion techniques. While they found that removal of the several pieces of potential vorticity (PV) associated with the midlatitude trough resulted in little reintensification of the tropical cyclone remnants, removal of the PV associated with the tropical cyclone resulted in extratropical cyclone development similar to that of Earl. However, it must be noted that their technique did not remove the moisture associated with the TC, just the balanced fields associated with the PV anomaly; thus, the moisture remnants of the tropical cyclone may have affected their results. Klein et al. (2002) attributed some aspects of the reintensification to midlatitude contributions but also found different contributions from the tropical cyclone. In such real-data case studies, it can be difficult to ascertain cause-and-effect relationships when studying ET. High-resolution idealized simulations of ET will be studied here using only the atmospheric portion of the Coupled Ocean–Atmosphere Mesoscale Prediction System (COAMPS; Hodur 1997). The advantage of simulations using such a modeling system is that the varying atmospheric conditions among cases can be controlled. Specifically, this study differs from those of Klein et al. (2002) and McTaggart-Cowan et al. (2001) in that it directly examines the effect of modifying the upper-level trough strength without changing either the tropical cyclone characteristics or the relative locations of the interacting weather systems. As noted in Klein et al. (2002), changing the relative location of the systems by only a small amount can dramatically affect the subsequent interaction. This idealized study only examines the effects of varying the strength of the midlatitude upper-level trough as an intermediate step to understanding the reintensification stage of the ET of a tropical cyclone.

The structure of the paper is as follows: A description of the modeling system is provided in section 2. The simulated development in the midlatitude environment in the absence of a tropical cyclone is examined in section 3 and validated against previous studies of midlatitude oceanic cyclogenesis. In section 4, some validations of the simulations of the interaction between the
midlatitude circulation and a tropical cyclone are provided. The intensification processes associated with the interaction are examined via comparisons with the midlatitude-only simulations of section 3. Some inferences from the simulations are compared with the Klein et al. (2002) interpretations to develop an improved understanding of the physical processes involved in the reintensification stage of ET. A summary and conclusions are provided in section 5.

2. Description of the modeling system

a. Model description

The COAMPS model employed in the study is described in detail by Hodur (1997). The system is non-hydrostatic, has multiple nested grids, and includes a Kain–Fritsch (KF; 1993) representation of convection, explicit moist physics, and boundary layer processes. The primitive equations are solved on a Lambert conformal grid, with a terrain-following \( \sigma \) coordinate in the vertical. The model has 36 layers from \( \sigma = 0 \) to 1, with the vertical boundaries at 30 km and the surface. Whereas vertical velocity is defined at the interfaces of the model layers, all other variables are carried at the midpoints of the layers. The horizontal grid has an Arakawa–Lamb C-staggering of the momentum variables \( (u \text{ and } v) \) with respect to the other variables.

In these simulations, the coarse and fine meshes have grid spacings of 81 and 27 km, respectively. The coarse domain of \( 87 \times 93 \) grid points (Fig. 1) is large enough to allow an adequate representation of the upper-level midlatitude trough development during the integration. The fine mesh of \( 124 \times 190 \) grid points captures the primary structural modifications of the storm as it interacts with the upper-level trough. The coarse grid supplies boundary values to the fine mesh, which in turn feeds information back to the coarse grid after the fine-mesh grid integration step is completed. The two-way interaction ensures that the fine-mesh structure is well represented on the coarse mesh. Consequently, only coarse-mesh figures will be presented here. Lateral boundary conditions (Perkey and Krietzberg 1976) around the coarse domain force the model-predicted variables near the outer boundary to adjust to the fixed initial values. The KF scheme is employed to treat subgrid-scale convective processes, while the explicit moist physics (Rutledge and Hobbs 1983) treats any grid-scale saturation. Additional parameterizations include subgrid-scale mixing (Deardorff 1980), surface fluxes (Louis 1979), and radiation (Harshvardhan et al. 1987).

b. Experimental setup

The initial environmental-wind vertical structure is based on composite data of western North Pacific ET cases (Ritchie and Elsberry 2001, step 3 of their Fig. 3). The horizontal variation in the environmental wind is based on a Gaussian distribution that results in a jet in the wind field near 47°N (not shown). Given this
In the case of the weak upper-level trough, very little perturbation that is added to the prescribed zonal flow. The time from the start of the intensification to achieve $P_{\text{max}}$ is also given.

<table>
<thead>
<tr>
<th>Upper-level trough</th>
<th>$V_{\text{max}}$ (m s$^{-1}$)</th>
<th>$P_{\text{max}}$ (mb)</th>
<th>$V_{\text{max}}$ (m s$^{-1}$)</th>
<th>Time to $P_{\text{max}}$ (h)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Weak</td>
<td>15</td>
<td>1003</td>
<td>11</td>
<td>80</td>
</tr>
<tr>
<td>Moderate</td>
<td>25</td>
<td>991</td>
<td>17</td>
<td>112</td>
</tr>
<tr>
<td>Strong</td>
<td>35</td>
<td>977</td>
<td>22</td>
<td>79</td>
</tr>
</tbody>
</table>

wind structure, the corresponding mass and temperature fields are derived to be in geostrophic and hydrostatic balance, with the mean thermal sounding specified as the composite prefrontal depression sounding of McBride and Zehr (1981). The entire domain is over ocean. A time-invariant sea surface temperature gradient is specified to match the near-surface air temperature gradient so that surface fluxes of moisture and heat do not erode the near-surface temperature structure through the simulation. The composite vertical profile of relative humidity (McBride and Zehr 1981) is specified everywhere so that the meridional temperature gradient implies moist (dry) air to the south (north). A balanced upper-level potential vorticity perturbation with a Gaussian vertical wind structure maximum at 380 mb is then added to this environment to represent the midlatitude upper-level trough. The maximum wind associated with the weak-trough perturbation is 15 m s$^{-1}$, and that of the strong trough is 35 m s$^{-1}$.

The initialization of the tropical cyclone is the same as in Ritchie and Elsberry (2001). The tropical cyclone is spun up in a quiescent environment until an approximately steady-state intensity of 960 mb is reached and the cloud fields are fully developed. This initial tropical cyclone has a maximum wind of 55 m s$^{-1}$ at a radius of 70 km. The core structure is approximately symmetric, with cyclonic winds extending to 50 mb in the core and a maximum temperature anomaly of 12 K at 450 mb (see Fig. 5 of Ritchie and Elsberry 2001). The 300-mb winds turn weakly anticyclonic at a 400-km radius.

All simulations are integrated until either filling of the extratropical cyclone begins (e.g., Figs. 3 and 9) or the system moves out of range of the fine mesh (120–144 h).

### 3. Cyclone development in the midlatitude environment only

The simulations described in Table 1 are of midlatitudinal cyclogenesis in association with the three upper-level troughs in the absence of a tropical cyclone. The initial strength of the upper-level trough is varied from weak (Fig. 2a) through moderate (not shown) to strong (Fig. 2b) based on the strength of the potential vorticity perturbation that is added to the prescribed zonal flow. In the case of the weak upper-level trough, very little surface cyclone development occurred during the simulation (Table 1), although the system may have continued to develop downstream of the region of interest. For the other two cases, a surface minimum pressure was achieved during the integration period, with the deepest development (977 mb and 22 m s$^{-1}$) in the case of the strong upper-level trough (Fig. 3). The development of the low-level (850 mb) equivalent potential temperature ($\theta_e$) and selected 6-h precipitation values are shown in Fig. 4 for the strong-trough case. Note in Figs. 4a and 4c the densely packed contours of $\theta_e$ that indicate development of strong cold and warm fronts. By 72 h of simulation (Fig. 4c), the $\theta_e$ contours indicate that the cyclone is becoming occluded, and it reaches its strongest intensity of 977 mb by 79 h of simulation, before beginning to fill. The 6-h precipitation patterns at 60 and 72 h are similar to that expected during a midlatitude cyclogenesis event (Petterssen and Smebye 1971). The maximum precipitation is located in the northeastern quadrant in association with the warm front. The cyclonic flow induced by the developing surface low pressure system forces the moist surface air up the steeply sloping isentropes in the northeast quadrant (not shown). The strongest flow is at the location of the steepest gradient of isentropes, and it is just to the north of this location that the maximum precipitation occurs. Rather than triggering deep convection, a stable, stratiform precipitation [defined as rain rates < 25 mm (3h)$^{-1}$] develops as shown in Fig. 4b and, to a lesser extent, in Fig. 4d.

The development of the 500-mb height and surface pressure fields for the strong-trough case is shown in Fig. 5 at 36, 60, and 84 h. Notice the building of the ridge downstream of the trough by 60 h, and the development of a secondary trough such that a wavenumber-2 pattern is present across the coarse domain. Particularly at 60 h of simulation (Fig. 5b), the primary trough has a marked tilt with height to the northwest that is indicated by the locations of the surface pressure minimum and 500-mb height minimum. The surface low pressure center is located in the left exit region of the upper-level jet maximum and associated upper-level divergence (not shown). By 84 h (Fig. 5c), after the strongest intensity has been achieved (Fig. 3), the system has become vertically stacked, as expected.

In contrast to the development of the strong-trough case, the weak-trough case has very little development by 72 h (Fig. 6). The 500-mb heights (Fig. 6a) indicate a weakly amplified wavenumber-1 pattern in the coarse domain with only a small ridging downstream of the trough. Weak secondary surface pressure centers have developed behind the primary low along the baroclinic zone (not shown). One major difference from the strong-trough case is a less organized and weaker development of a low-level equivalent potential temperature gradient for the weak case (cf. Figs. 6b, 4). Less upslope flow of moist southerly air occurs in the northeast sector of the storm, so less cloudiness and precipitation is sim-
ulated (not shown). In addition, the upper-level winds (not shown) in the jet region of the midlatitude circulation do not amplify as in the strong-trough case, and the associated upper-level divergence is considerably weaker throughout the simulation (not shown).

The Klein et al. (2002) definition of the development potential is shown in Fig. 7 for all trough cases. Note that all three Petterssen development processes are contributing to the strong development potential region to the northeast of the developing surface cyclone in the strong-trough case (Fig. 7a, dark gray shading). This strong development potential region is just in advance of the surface cyclone, which Klein et al. (2002) also found for real cases of surface development. At this time, the surface center has a minimum pressure of 998 mb and is intensifying (Fig. 3). The contributions to the development potential region are primarily associated with low-level warm temperature advection and midlevel vorticity advection, with a small region of upper-level divergence. By 72 h of the strong-trough simulation, the development potential region has broadened and weakened. In addition, much of the region of maximum development potential has detached from the surface cyclone and is farther east (Fig. 7b). At this time, the cyclone is becoming occluded (recall the precipitation and surface equivalent potential temperature patterns in Fig. 4). Because the region of strongest development potential has weakened and become detached from the surface cyclone at the time when the cyclone is no longer intensifying, the development potential may be a useful indicator that development is ending. Klein et al. (2002) found a similar evolution of the development potential in their real-data predictions.

For the weak case, a smaller and weaker region of development potential is indicated (Figs. 7c,d). The contributions to the development potential region are from a weaker low-level temperature advection and midlevel vorticity advection compared with the strong-trough
case, with no contribution from an upper-level divergence that meets the threshold requirements. A better phasing with the cyclone center is predicted at 48 h with an overlap of the two development variables, which is indicated in Fig. 7d by a small region of medium gray shading.

A similar, but weaker, development potential pattern is simulated for the moderate-trough case (Figs. 7e,f) than for the strong-trough case. As would be expected, the deepening rate is correspondingly less in the moderate-trough case (Fig. 3). The moderate-trough case highlights how dependent the surface cyclone development is on superposition, with a development potential region having at least two contributing factors. Very little development potential is indicated at 12 h (Fig. 7e), and the corresponding sea level pressure field (Fig. 8a) has varied little from the initial conditions (not shown). In addition, the rate of deepening of the surface pressure is small (Fig. 3). By 60 h, two regions of development potential are simulated, with the easternmost region having the greatest extent and strength (Fig. 7f).

The corresponding sea level pressure field (Fig. 8b) indicates that the strongest surface cyclone development is occurring in the easternmost area. In addition, the rate of intensification is substantially increased from earlier in the integration (Fig. 3). As in the strong and weak cases described above, the development potential parameter of Klein et al. (2002) does appear to be a reasonable indicator of the location and the magnitude of cyclone development that may occur.

If the characteristics of the midlatitude circulation are the main factor in determining the strength of the extratropical intensification of the poleward-moving tropical cyclone remnants, then these simulations imply that an interaction with a weak upper-level trough during ET will not have the potential to reintensify as much as during an interaction with a moderate or strong upper-level trough. This hypothesis is investigated in the next series of simulations.

4. Simulated interactions of an upper-level trough with a tropical cyclone

a. Development of the surface cyclone and trends in intensification

The characteristics of the surface cyclone for the three simulations of weak, moderate, and strong troughs, as in section 3 but with a TC inserted at \( t = 0 \) in the same location 15°S, 25°E of the upper-level trough, are summarized in Table 2 and Fig. 9.

The primary difference between the trough-only and the trough-with-TC simulations is that the trough-with-TC simulations intensify to a much greater depth during a similar time frame compared with the trough-only

<table>
<thead>
<tr>
<th>Midlatitude trough</th>
<th>( P_{\text{init}} ) (mb)</th>
<th>( P_{\text{min}} ) (mb)</th>
<th>( \Delta P_{\text{init}} ) (mb)</th>
<th>( V_{\text{max}} ) (m s(^{-1}))</th>
<th>Time (h)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Weak</td>
<td>998</td>
<td>967</td>
<td>31</td>
<td>27.9</td>
<td>43 (16)</td>
</tr>
<tr>
<td>Moderate</td>
<td>991</td>
<td>965</td>
<td>26</td>
<td>29.4</td>
<td>55 (10)</td>
</tr>
<tr>
<td>Strong</td>
<td>987</td>
<td>959</td>
<td>28</td>
<td>27</td>
<td>38 (13)</td>
</tr>
</tbody>
</table>
Fig. 5. Simulation of a strong trough with predicted 500-mb geopotential heights (m; contours every 60 m) superposed on sea level pressure (shaded every 4 mb for values ≥1000 mb) at (a) 36, (b) 60, and (c) 84 h. The location of the surface minimum pressure is indicated (⊙).

Fig. 6. Simulation of a weak trough at 72 h: (a) 500-mb geopotential heights (m; contours every 60 m) and (b) 850-mb equivalent potential temperature (4-K contours). The location of the surface minimum pressure center is indicated (⊙).

Fig. 9 and Table 2. First, progressively higher surface pressures and lower wind speeds are attained by the tropical cyclone during the transformation stage for progressively weaker strengths of the upper-level trough. This is because the deep-layer-mean steering flow over the tropical cyclone is weaker (stronger) in the weak- (strong-) trough case. However, the westerly shear generated by the midlatitude circulation during transformation is approximately the same in each case. This is illustrated in Fig. 10. During the transformation stage, the westerly shear associated with both cases steadily increases to 10–12 m s\(^{-1}\). However, the deep-layer-mean steering flow is much stronger for the strong upper-level-trough than for the weak upper-level-trough case. Thus, the tropical cyclone associated with the weak upper-level trough moves into the favorable development potential region generated by the midlatitude circulation by 90 h of simulation, compared with 60 h for the tropical cyclone associated with the strong upper-level trough. Consequently, the weak-trough tropical cyclone spends more time in the strong westerly shear regime associated with the transformation stage (e.g., Ritchie and Elsberry 2001; Klein et al. 2000), becomes much more tilted and sheared, and weakens considerably more before reintensification begins. This is similar to the findings of McTaggart-Cowan et al. (2001) for their weaker-trough case during the transformation process, which was impacted by much stronger shear than was their control. The main difference between our simulations and theirs is that the phase lock between the upper-level trough
and tropical cyclone in their weak upper-level-trough case is altered compared with their control so that not so much reintensification occurs, while ours is not.

Another result from Fig. 9 and Table 2 is that the final intensity attained by the extratropical cyclone in the weak- and moderate-trough cases is not correlated with the relative strength of the midlatitude trough since both cases attained similar intensities. It seems likely that the maximum intensity that can be achieved during the reintensification stage of ET is more dependent on the large-scale characteristics of the midlatitude environment than on the details of the upper-level-trough structure. For the weak and moderate simulations, the final intensity of the extratropical cyclone was similar, probably because the general baroclinic-environment structure was similar. The strong-trough case attained not only a slightly lower minimum pressure but also a slightly lower surface maximum wind speed for the same baroclinic environment. This deeper sea level pressure may be due in part to the more poleward movement of the strong extratropical cyclone, and the momentum measurement may be a truer measure of actual intensity.

A third result in Fig. 9 and Table 2 is that the change in pressure from maximum surface pressure at transformation to minimum surface pressure during reintensification in the weak trough-only case and moderate trough-only case is not simply related to the relative strength of the midlatitude trough. Whereas the deepening for the weak trough-only case was less than that for the moderate trough-only case, the weak trough-only case deepened more than the moderate trough-only case.

A fourth result from the weak- and strong-trough comparison is an apparent difference in translation speeds during the early reintensification period. Whereas the motion of the surface minimum pressure associated with the weak trough-only case is approximately constant with time (Fig. 11b), a discontinuity in translation occurs between 72 and 96 h for the moderate trough-only case (not shown) and between 48 and 72 h for the strong trough-only case (Fig. 11b). In these two cases, the tropical cyclone remnants appear to accelerate rapidly to the north, but what actually occurs is that the tropical cyclone remnants have dissipated, and it is the surface circulation associated with the upper-level trough farther north that is developing and becomes the dominant cen-
Fig. 8. Simulation of a moderate trough with predicted 500-mb geopotential heights (m; contours every 60 m) superposed on sea level pressure (shaded every 4 mb for values ≤1000 mb) at (a) 12 and (b) 60 h. The location of the surface minimum pressure is indicated (◇).

In summary, whereas the simulated weakening of the tropical cyclone that occurs during the transformation stage relates well to the strength of the upper-level trough, the magnitude of the development during the reintensification stage does not. All three troughs developed deep, similar-strength cyclones (as measured by wind field). In addition, the most intensification and longest period of rapid intensification were associated with the weakest trough. Possible reasons for these intensification differences will be examined in the following sections, along with the intensification mechanisms for the weak- and strong-trough cases.

b. Interaction between a tropical cyclone and a weak midlatitude trough

The interaction between a tropical cyclone and weak midlatitude trough represents the most extreme contrast of development when compared with the no-TC control simulation. In the comparable control, almost no development was associated with the weak midlatitude trough. Interestingly, when a transforming tropical cyclone was properly phased with the weak trough, the largest reintensification (31 mb and 9 m s\(^{-1}\)) of the tropical cyclone remnants resulted with the longest period of rapid intensification (Fig. 9; Table 2).

Initially, the midlatitude trough and the tropical cyclone are distinct features in the 500-mb height and surface pressure fields (Fig. 12a). Prior to 84 h of integration, the tropical cyclone undergoes transformation and weakens considerably more than the other simulations as it interacts with the increasing midlatitude shear (Figs. 9, 12a,b). Subsequent to 84 h of integration, the system begins to intensify (Figs. 12c,d), and during this time the system is tilted westerly with a height similar to many documented cases of ET (e.g., DiMego and Bosart 1982; Klein et al. 2002; Sinclair 1993; Thorncroft and Jones 2000).

One reason the trough-with-TC simulation intensifies so much more than the trough-only simulation is the interaction between the tropical cyclone and low-level baroclinic zone. During the transformation stage of extratropical transition, the low-level temperature and moisture fields associated with an essentially zonal baroclinic zone are distorted by the circulation associated with the tropical cyclone (Schultz et al. 1998; Klein et al. 2000; Ritchie and Elsberry 2001) such that warm advection occurs to the east of the tropical cyclone and cold advection to the west. This results in a much stronger temperature and moisture gradient to the northeast of the tropical cyclone than occurs when no tropical cyclone is present (Figs. 13a, 6b). In addition, an injection of higher-\(\theta_e\) air in the lower troposphere associated with the tropical cyclone remnants is also absorbed into the developing extratropical cyclone (Fig. 13a). Thus, a much stronger surface warm front develops to the northeast of the developing extratropical cyclone than occurs in the trough-only case (not shown). The lifting of this moist air results in precipitation that is greater in both amount and areal extent (Fig. 13b). By 120 h, the extratropical storm has deepened to 970 mb when the tropical cyclone is present compared with 1003 mb in the trough-only simulation. However, the system has begun to be vertically stacked and become occluded, as indicated by the low-level \(\theta_e\) and precipitation fields (Figs. 13c,d). Thus, the extreme rate of intensification has ended (Fig. 9). By 132 h, the system has become vertically stacked (Fig. 12d) and has achieved minimum sea level pressure (967 mb). Subsequent to this time, it begins to slowly weaken (Fig. 9).

Another reason the cyclone in the simulation including the tropical cyclone intensifies so much more than the trough-only simulation is substantial differences in the upper-level fields. At 36 h, the trough-only (Fig. 14a) and the trough-with-TC (Fig. 14c) cases have similar upper-level patterns, and the jet maxima have comparable strengths and are at about the same relative lo-
cations. However, the tropical cyclone center is still more than 1000 km south of the upper-level jet maximum. By 60 h, the maximum wind speed in the trough-only case (Fig. 14b) is about 2500 km east of the surface cyclone so that any development potential associated with divergence in the upper-level jet maximum is far away. However, the maximum wind speed in the tropical cyclone case (Fig. 14d) is just east of the development.

**Fig. 9.** Time series for each of the three simulations of an interaction between a midlatitude trough and a tropical cyclone: (a) minimum surface pressure and (b) azimuthally averaged maximum wind speed.

**Fig. 10.** Deep-layer mean flow and vertical wind shear (200–850 mb) averaged over a 900-km² box centered on the tropical cyclone for the weak and strong trough-with-TC cases.
FIG. 11. Geopotential height (m) field at 500 mb for the initial time for (a) the weak trough with a TC and (b) the strong trough with a TC. The thick solid line is the track of the surface minimum pressure over the 120-h period.

FIG. 12. Predicted 500-mb heights (contours every 60 m) and surface pressure (shaded every 4 mb for values ≤1000 mb) during the interaction between a tropical cyclone and a weak midlatitude trough at (a) 24, (b) 84, (c) 108, and (d) 132 h. The surface minimum in pressure is indicated (○).

region and has increased more than 15 m s$^{-1}$ compared with the trough-only case. This wind speed increase is due to the warm upper-tropospheric outflow associated with the tropical cyclone interacting with the polar jet stream (e.g., DiMego and Bosart 1982; Klein et al. 2000; Sinclair 2002) such that the horizontal temperature gradient below the jet increases and the height of the adjacent tropopause increases. As the wind speed in the jet maximum increases, the associated upper-level divergence increases (Fig. 15). Thus, upper-level divergence is considerably enhanced in the development region by the outflow from a tropical cyclone compared to the simulation with no tropical cyclone present (cf. Figs. 15b,d).

The key to why the cyclone in the weak-trough case intensifies more rapidly than that in the moderate-trough case (Fig. 9) appears to lie in small differences in the parameters that contribute to the development potential region. That is, small differences in the temperature advection, upper-level divergence, and midlevel PVA appear to result in differences in the intensification rate of the extratropical cyclone. We have not tested whether the absolute values of the contributing parameters are also important (and not just that they exceed their threshold values). In addition, the length of time that the cyclone remains in the prime development potential region may also be important. For a given available potential energy (APE), the amount that can be converted to kinetic energy (KE) determines the intensification of the surface cyclone. If the system can remain in the conversion zone long enough to convert the entire APE to KE, then maximum intensification may occur. Based on the initial environmental structure, it may be that both of the simulated cyclones remained within the prime
development region long enough to intensify to the minimum central pressure possible given the environmental conditions. These simulations support the hypothesis of Klein et al. (2002) that the structure of the midlatitude environment may be dictating the minimum intensity that the reintensifying cyclone can achieve.

c. Interaction between a tropical cyclone and a strong midlatitude trough

The strong midlatitude-trough-case simulation is similar to a class of extratropical transitions defined originally by Matuno and Sekioka (1971) as a “compound transition,” and more recently by Foley and Hanstrum (1994) and Klein (2000) as a “merger” between the tropical cyclone and surface cyclone associated with the upper-level trough. The climatology developed by Klein (2000) yielded only four mergers from a total of 30 cases, so the strong midlatitude-trough case is similar to a type of ET that only rarely occurs in nature. In this case, the midlatitude trough is so strong that a surface cyclone develops in association with the trough before the transforming tropical cyclone reaches the development region. As the transforming tropical cyclone remnants advect into the vicinity of the surface pressure center associated with the upper-level trough, a rotation and merger of the two surface centers occurs (Fig. 16), similar to that observed by Klein (2000) for Typhoon Joy. Interestingly, McTaggart-Cowan et al. (2001) also reported secondary surface low developments for their stronger upper-level-trough case, although whether the same mechanisms reported here had a direct impact on their results is unclear.

Initially, the midlatitude trough and the tropical cyclone are distinct features in the 500-mb height and surface pressure fields (Fig. 16a). Prior to 72 h of integration, the development of the surface pressure center associated with the midlatitude trough is quite similar to the development in the strong trough-only case (cf. Figs. 5c, 16b). At the same time, the tropical cyclone weakens (Fig. 16b) as it begins transformation in the midlatitude westerly shear. However, between 60 and ~78 h of simulation, the tropical cyclone approaches the midlatitude-trough center such that absorption of the TC remnants by the midlatitude-trough center occurs (Figs. 16b–e). The remnant of the tropical cyclone can be seen in Fig. 16d as a southerly lobe in the surface pressure pattern, and 6 h later (Fig. 16e) the lobe has almost completely disappeared. At about this time, the system begins a 14-h period of rapid intensification, which is consistent with the westward tilt with height that also occurred in the strong trough-only simulation. By 84 h, the extratropical storm pressure is already down to 964 mb compared with 977 mb in the trough-only simulation. However, by 96 h the system has become vertically stacked (Fig. 16f) and minimum sea level pressure is achieved (959 mb and 27 m s\(^{-1}\)). One difference between this simulation and the cases identified by Klein (2000) as “mergers” is that the TC remnants are absorbed by the trough circulation. Klein (2000) observed in three out of four cases that the tropical cyclone remnants remained strong enough to absorb the trough circulation rather than the other way around.

Similar to the weak trough-with-TC simulation, one reason the surface low center deepens more in the strong trough-with-TC case than in the trough-only case is the slightly higher \(\theta_e\) values in the lower troposphere associated with the tropical cyclone circulation (cf. Figs.
17a,b and 4a). In addition to higher-\( \theta_e \) air associated with the core of the tropical cyclone remnants that is absorbed directly into the upper-level trough (Figs. 17b,c), higher-\( \theta_e \) air from the Tropics is advected around the east of the tropical cyclone remnants by the tropical cyclone outer circulation. This warm, moist air pushes into the colder air associated with the baroclinic zone and enhances the low-level thermal gradient (Klein et al. 2000) and then, lifting over the tilted isentropes, enhances precipitation and latent heat release in the northeast quadrant of the developing cyclone (cf. Figs. 17d, 4d).

An additional reason for greater deepening in the strong trough-with-TC case lies in the upper-level jet maximum, which is enhanced similarly to that for the weak-trough case illustrated in Fig. 14. At 36 h, when the cyclone is still more than 1000 km south of the jet maximum, the upper-level patterns between the with-
Fig. 16. Predicted 500-mb heights (contours every 60 m) and surface pressure (shaded every 4 mb for values < 1000 mb) during the interaction between a tropical cyclone and a very strong midlatitude trough at (a) 24, (b) 60, (c) 66, (d) 72, (e) 78, and (f) 96 h. The surface minimum pressure centers associated with the tropical cyclone and midlatitude trough are indicated (†).

Fig. 17. Predicted 850-mb equivalent potential temperatures (contours every 4 K) during the interaction between a tropical cyclone and a strong upper-level trough: (a) 48-, (b) 60-, (c) 72-, and (d) 6-h precipitation (1, 5, 10, and 50 mm/6 h) corresponding to (c). The location of the surface minimum pressure center is indicated (†).
TC (Fig. 18a) and no-TC cases (Fig. 18c) are very similar. By 60 h, the outflow from the tropical cyclone has started to interact with the upper-level jet (Fig. 18b), and a stronger wind speed maximum with increased upper-level divergence (not shown) is simulated compared with the no-TC case (Fig. 18d).

Because of the merging of the two systems, the development potential region in the with-TC simulation is larger in both areal extent and strength (Fig. 19) compared with the control. Little difference exists for the development potential region at 24 h (cf. Figs. 19a, 7a). By 72 h, as the transforming tropical cyclone is merging with the midlatitude cyclone, both the strength and the areal extent of the development potential region is increased (Fig. 19b) compared with the control (Fig. 7b). Beginning approximately 60 h of integration, the tropical cyclone outflow interacting with the jet stream mainly contributes upper-level divergence to the development potential region (cf. Figs. 18b,d with 16b,d) that is already consisting mostly of strong low-level
temperature advection and PVA from the midlevel trough.

The areal and magnitude increase in the development potential region in the strong trough-with-TC case (Fig. 19b) coincides with the start of the reintensification period of the strong trough-with-TC case (Fig. 9). The merging of the two systems then occurs between 60 and 78 h of integration (Figs. 16b–e), and at 71 h the period of rapid intensification begins (Table 2). The period of rapid intensification continues until 84 h when the strong trough-with-TC begins to show signs of occlusion (Fig. 19c), and the cyclone reaches its maximum intensity at 96 h (Fig. 9). The time of maximum intensity of the strong trough-with-TC case coincides with the dissipation of the development potential region (Fig. 19d) due to weakening of low-level temperature advection and reduction in both the upper-level divergence and midlevel PVA. That is, as the cyclone becomes vertically stacked (e.g., Fig. 16f), the development terms in the Petterssen equation are reduced (Fig. 19d).

5. Summary and conclusions

Six simulations are used to describe the development processes of an extratropical cyclone during the reintensification stage of ET. Since the objective is to elucidate the role of the strength of the midlatitude upper-level trough, the first set of three simulations (the controls) examined the extratropical cyclogenesis associated with upper-level troughs that are characterized as weak, moderate, and strong. The second set of three simulations, the interactions of a tropical cyclone with each of the three upper-level troughs, is compared with the control simulations to illustrate the contributions to the extratropical transition of the tropical cyclone.

As expected, the deepest surface cyclone (977 mb) developed in association with the strongest midlatitude trough, and the weakest surface cyclone (1003 mb) developed in association with the weakest midlatitude trough. This expected trend is consistent with the hypothesis of Klein et al. (2002) that the development of the surface cyclone as part of the reintensification stage of the ET of a tropical cyclone should depend strongly on the characteristics of the midlatitude circulation into which the TC is moving.

Whereas the second set of simulations including the TC all resulted in greater deepening, the complexity of the resulting cyclogenesis events indicates that the reintensification stage depends on more than just the strength of the midlatitude upper-level trough.

First, the duration of the transformation stage of the tropical cyclone is related to the strength of the midlatitude trough prior to the beginning of the reintensification. That is, the transformation stage lasted longer and resulted in weaker tropical cyclone remnants for the weak-trough case compared with the strong-trough case, and the resulting tropical cyclone intensity with the moderate trough is intermediate, as might be expected.

The reason for the weaker (stronger) tropical cyclone at the end of the transformation stage is that a weaker (stronger) northeastward steering flow is associated with the weak- (strong-) trough environment, which results in the TC remaining in the hostile vertical wind shear environment for a longer (shorter) period until a favorable phasing with the midlatitude trough was achieved such that reintensification occurred.

Second, the final intensity of the extratropical cyclone does not appear to be dependent only on the strength of the midlatitude trough. In all three trough-with-TC cases, the final intensities achieved were almost identical (967 mb and 27.9 m s$^{-1}$, 965 mb and 29.4 m s$^{-1}$, 959 mb and 27 m s$^{-1}$), especially when compared with the trough-only simulations (1003 mb and 10.9 m s$^{-1}$, 991 mb and 17 m s$^{-1}$, and 977 mb and 22.1 m s$^{-1}$). An important factor is the phasing of the tropical cyclone with the midlatitude trough that resulted in substantial enhancement of the upper-level divergence, as discussed by DiMego and Bosart (1982) and Klein et al. (2002). Based on these simulations, an important conclusion is that a weak midlatitude trough interacting with tropical cyclone remnants may have as much potential to intensify as does a moderate or strong trough, and may have longer periods of rapid intensification.

Finally, the development potential parameter that was defined by Klein et al. (2002) based on the three main factors in the Petterssen development equation (upper-level divergence, midlevel PVA, and low-level temperature advection) appears to have utility in predicting where and whether extratropical cyclogenesis will occur. In future work, these simulations are extended to examine further how the phasing of the tropical cyclone with the three midlatitude circulation strengths also affects the subsequent reintensification of the tropical cyclone remnants.

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