The Downstream Extratropical Flow Response
to Recurving Western North Pacific Tropical Cyclones

by

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1. Introduction

a. Motivation and purpose

Tropical cyclones (TCs) that recurve into the midlatitudes while undergoing extratropical transition (ET) have long been recognized as hazards due to their capacity to produce high winds, heavy rain, and storm surge (e.g., Pierce 1939; Namias 1963). Perhaps less appreciated is that recurving TCs also may lead to high-impact weather thousands of kilometers downstream. By perturbing the extratropical jet stream, a recurving TC may induce Rossby wave amplification and dispersion (e.g., Riemer et al. 2008; Harr and Dea 2009; Riemer and Jones 2010). Frequent TC recurvatures over the western North Pacific (WNP) are of particular concern to interests in North America because Rossby wave dispersion from the North Pacific has been linked to the onset of large-scale flow anomalies and high-impact weather events over North America (e.g., Archambault et al. 2010).

Individual case studies reveal that the downstream extratropical flow response to recurving TCs varies considerably (e.g., Harr and Dea 2009). The response can range from the excitation or amplification of a high-amplitude Rossby wave train (RWT hereafter), to the intensification and downstream extension of the jet stream, to little discernable flow response. The combination of factors governing this difference in behavior is not well understood, although characteristics of the large-scale flow pattern, the TC itself, and the phasing of the TC with the extratropical flow during ET are believed to be important (e.g., Reynolds et al. 2009; Harr and Dea 2009; Riemer and Jones 2010).

An issue related to the range of possible extratropical flow responses to recurving WNP TCs is that recurving TC episodes are frequently associated with reduced predictability, which is manifested as increased model forecast error and/or increased spread among members of an ensemble prediction system (EPS). For example, large errors in model forecasts of the midtropospheric flow over the North Pacific have been observed during episodes of recurving and transitioning WNP TCs (Jones et al. 2003). The reduced predictability that arises during recurving TC episodes may be due to the high sensitivity of ET to the phasing of the TC with the extratropical flow (e.g., Klein et al. 2000; Jones et al. 2003; Ritchie and Elsberry 2007; Anwender et al. 2010). Reduced predictability may also arise from the upscale growth of error and uncertainty associated with diabatic ridge amplification accompanying ET. In this situation, forecast error and uncertainty may spread well downstream in conjunction with Rossby wave dispersion (e.g., Reynolds et al. 2009; Anwender et al. 2010).

The overarching goal of this dissertation is to understand the factors that modulate the observed downstream extratropical flow response to recurving and transitioning WNP TCs. The following questions will be addressed: (i) How do characteristics of the large-scale flow pattern, the TC itself, and the phasing of the TC with the extratropical flow (e.g., the TC–jet stream interaction) modulate the extratropical flow response to recurving WNP TCs? (ii) What are the important synoptic–dynamic processes accompanying high-amplitude extratropical flow responses and high-impact weather events associated with recurving WNP TCs? (iii) What factors influence the predictability of recurving WNP TC episodes?

b. Organization of dissertation

The dissertation is organized as follows: The remainder of Chapter 1 contains a literature review. A climatology of recurving WNP TCs is presented in Chapter 2, whereas Chapter 3 contains the results of composite analyses constructed using cases identified in the climatology. Chapter 4 contains two case studies of recurving WNP TC episodes associated with strong TC–jet stream interactions, pronounced flow reconfigurations and high-impact weather over North America, and, in the case of one recurving WNP TC episode, reduced predictability. A discussion of the key findings is found in Chapter 5. Finally, Chapter 6 contains a summary and potential future research paths.

c. Literature review

i. General characteristics of TC recurvature

In an otherwise quiescent environment, a westward-moving TC embedded in Northern Hemispheric (NH) tropical easterlies will eventually turn toward the northwest due to beta drift [i.e., steering induced by a beta gyre that forms via planetary vorticity advection induced by the cyclonic circulation of the TC (e.g., Holland 1983, 1984; Carr and Elsberry 1990)]. If the TC is located on the southern edge of a potential vorticity (PV) strip, which is a zonally elongated area of cyclonic PV generated by diabatic heating associated with intertropical convergence zone convection, its turn toward the northwest will tend to be more marked (e.g., Evans et al. 1991; Ferreira and Schubert 1997). In the latter scenario, relative vorticity advection by the TC circulation acts to reinforce planetary vorticity advection by the TC circulation, inducing a more northward

TC motion than would result from beta drift alone. In addition, a westward-moving TC may turn northwestward in response to southerly flow occurring along the western periphery of a subtropical anticyclone (e.g., Camargo et al. 2007) or induced by an equatorially trapped wave such as an equatorial Rossby or mixed Rossby–gravity wave (e.g., Matsuno 1966).

Once a TC tracks northwestward into higher latitudes, it may undergo recurvature (i.e., a further heading change from northwestward to northeastward) as it encounters upper-tropospheric westerlies. During and following recurvature, the structure of the TC typically acquires extratropical characteristics (i.e., the TC undergoes ET) as it becomes embedded in the upper-tropospheric westerlies.

The environmental factors governing TC recurvature are well known. An early study by Riehl and Shafer (1944) documents that TCs tend to recurve in response to upper-tropospheric westerlies associated with the equatorward penetration of an extratropical trough. Subsequent studies establish techniques for forecasting TC track, speed, and acceleration in the event of recurvature or its possibility, all which make use of the concept that TCs tend to be steered by the mid-to-upper-tropospheric flow (e.g., George and Gray 1976, 1977; Chan and Gray 1982; Cheng-Lan and Sadler 1983; Hodanish and Gray 1993).

From a climatological perspective, WNP TC recurvature is favored under certain large-scale flow conditions. During WNP TC recurvature episodes, anomalous mid-tropospheric cyclonic and anticyclonic circulations tend to be present over the extratropical WNP and South China Sea, respectively (e.g., Harr and Elsberry 1991; Camargo et al. 2007). In addition, WNP TC recurvature episodes are more frequent during the warm phase of the El Niño–Southern Oscillation (ENSO; e.g., Bjerknes 1969) and less frequent during the cool phase of ENSO (e.g., Wang and Chan 2002; Elsner and Liu 2003). More frequent WNP TC recurvature episodes during the warm ENSO phase (El Niño) are attributed to enhanced upper-tropospheric southeasterly steering flow around 135°E associated with a deeper-than-normal east Asian upper-tropospheric trough. Conversely, less frequent WNP TC recurvature episodes during the cool ENSO phase (La Niña) are attributed to reduced upper-tropospheric southeasterly steering flow around 135°E associated with a weaker-than-normal time-mean upper-tropospheric trough over eastern Asia (Wang and Chan 2002).

In addition, the tendency for WNP TCs to form farther eastward during El Niño and farther westward during La Niña with respect to climatology (e.g., Chan 1985; Wang and Chan 2002) also is found to impact WNP recurvature frequency. The characteristic eastward shift in the WNP TC genesis region during El Niño is associated with an increased likelihood that a westward tracking WNP TC will encounter an upper-tropospheric trough and recurve (Elsner and Liu 2003). Conversely, the characteristic westward shift in the WNP TC genesis region during La Niña is associated with a decreased likelihood that a westward tracking WNP TC will encounter an upper-tropospheric trough and recurve.

WNP TC recurvature episodes occur primarily between late spring and late fall (i.e., May–December), with a peak in late August and early September (e.g., Jones et al. 2003[1]). A secondary peak in WNP recurvature episodes in the second half of October is

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[1] Although Jones et al. (2003) specifically examine ET rather than TC recurvature, their ET climatology likely closely corresponds to a TC recurvature climatology since nearly all TC recurvatures are associated with ET, and nearly all ETs are associated with TC recurvatures.

noted in an early study by Burroughs and Brand (Fig. 1.1; Burroughs and Brand 1973). The time of year in which WNP TC recurvature episodes are most frequent corresponds closely to the time of year in which WNP TCs occur most frequently (e.g., Fig. 1.1; Burroughs and Brand 1973; Jones et al. 2003).

Characteristics of recurving WNP TCs are influenced by the relatively large seasonal variability in the extratropical flow pattern over the WNP. An examination of the monthly mean extratropical flow pattern over eastern Asia, the North Pacific, and North America for May–December (Fig. 1.2) reveals that the WNP jet stream and associated baroclinic zone are shifted equatorward and are considerably stronger in late fall and early winter than in summer. The WNP jet stream and baroclinic zone also are shifted equatorward and are slightly stronger in late spring relative to summer (Fig. 1.2). These seasonal changes in the WNP jet stream are consistent with the tendency for WNP TCs to recurve at lower latitudes (e.g., Fig. 1.3; Riehl 1972; Burroughs and Brand 1973), recurve at a higher frequency (e.g., Fig. 1.1), recurve more sharply, and accelerate more rapidly after recurvature in late spring, late fall, and early winter than in summer (e.g., Burroughs and Brand 1973).

A potentially important finding of Burroughs and Brand (1973) that is not necessarily related to the seasonal variability of the WNP extratropical flow pattern is that WNP TCs at recurvature tend to be larger and more intense in the fall (September–November) than in spring or summer. This relationship between the size and intensity of WNP TCs and the season may be relevant to understanding the seasonal variation in the extratropical flow response to recurving WNP TCs. A more general point of interest to be addressed in Chapter 2 is how the Burroughs and Brand (1973) 1945–1969 climatology of recurving WNP TCs would compare to a more modern climatology of recurving WNP TCs.

ii. Extratropical transition

ET is a fundamental aspect of TC recurvature. As TCs moves into the extratropics, they encounter significant environmental changes such as horizontal moisture and sea surface temperature (SST) gradients, lower SSTs, increased westerly flow, increased vertical wind shear (i.e., baroclinicity), and increased planetary vorticity (e.g., Jones et al. 2003). In response to these environmental changes, TCs acquire characteristics of baroclinic, extratropical cyclones (ECs) through a two-stage process of transformation and extratropical reintensification. This two-stage conceptual model for ET is based on studies by Klein et al. (2000, 2002)[2]. Their analysis of 30 WNP ET events occurring from June through October during 1994–1998 reveals that TC structural changes during ET proceed in a relatively similar manner.

During the first step of the transformation stage (Fig. 1.4, left column), the outer circulation of the cyclone begins to interact with an east–west-oriented baroclinic zone (middle and bottom panels) and associated vertical wind shear. On the west side of the cyclone circulation, equatorward advection of cool, dry air (middle and bottom panels) reduces clouds and precipitation (top panel). On the east side of the cyclone circulation, however, poleward advection of warm, moist air (middle and bottom panels) maintains clouds and heavy precipitation (top panel). The cyclone circulation remains

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[2] Earlier conceptual models of ET are presented in studies by Matano and Sekioka (1971) and Foley and Hanstrum (1994).

approximately axisymmetric (middle and bottom panels) and the inner warm-core circulation associated with the cyclone remains upright (bottom panel).

In the second step of the transformation stage (Fig. 1.4, middle column), the baroclinic zone rotates cyclonically to a southwest–northeast orientation in response to the cyclone circulation (middle and bottom panels). The equatorward advection of cool, dry air (middle and bottom panels) limits clouds and precipitation on the west and south sides of the cyclone circulation (top panel), whereas the poleward advection of warm, moist air (middle and bottom panels) supports heavy precipitation on the east and north sides of the cyclone circulation (top panel). The cyclone circulation becomes asymmetric (middle and bottom panels) and the inner warm core circulation begins to tilt downshear (bottom panel) in response to increasing vertical wind shear.

In the third and final step of the transformation stage (Fig. 1.4, right column), a baroclinic wave develops. Equatorward advection of cool, dry air (middle and bottom panels) continues to limit clouds and precipitation on the west and south sides of the cyclone circulation (top panel), with cloudiness and deep convection on the east and north sides of the cyclone circulation maintained by a low-level jet (middle and bottom panels). The cyclone circulation becomes more asymmetrical (middle and bottom panels) and predominantly cold core, although a weak low-level warm core persists over the surface cyclone center (bottom panel). In this third step of the transformation process, distinct warm and cold frontogenesis is observed to the northeast and southwest of the cyclone center, respectively (e.g., Harr and Elsberry 2000).

The ET transformation stage typically is associated with a substantial weakening of the cyclone. The transformation stage is considered complete when the cyclone center is located on the cold side of a preexisting extratropical baroclinic zone, which corresponds well to the time that operational forecasters consider a cyclone extratropical (Jones et al. 2003).

The transformation stage of ET also can be understood using the cyclone phase space (CPS) paradigm of Hart (2003). Based on an analysis of CPS parameters of North Atlantic ET events, Evans and Hart (2003) define ET onset time to be when the lower-tropospheric thickness field of a transitioning cyclone reaches a certain threshold of asymmetry. Evans and Hart (2003) note that this ET onset time corresponds well with the ET transformation time defined by Klein et al. (2000).

After the transformation stage of ET, a subset of cyclones reintensifies as ECs instead of continuing to weaken. This processes of extratropical reintensification is deemed by Klein et al. (2000, 2002) to be the second stage of ET. During the reintensification stage, fronts form in the presence of high baroclinicity and strong vertical wind shear, with the warm front typically more prominent than the cold front (e.g., Harr and Elsberry 2000). The cloud and precipitation fields become increasingly asymmetric. The cyclone acquires a vertical tilt and typically becomes a deep cold-core system, although a shallow warm core can sometimes be maintained during the reintensification stage (e.g., Browning et al. 1998; Thorncroft and Jones 2000; Jones et al. 2003). Klein et al. (2000, 2002) consider ET complete when reintensification ceases.

Numerous studies have investigated factors that promote extratropical reintensification. Klein et al. (2000, 2002) and others note that the reintensification stage[3] is associated with the phasing of a transitioning TC with an upper-tropospheric disturbance such as an upstream trough (e.g., DiMego and Bosart 1982a,b; Harr and Elsberry 2000; Harr et al. 2000; Sinclair 2004; Ritchie and Elsberry 2007; Kitabatake 2008) or PV anomaly (e.g., Browning et al. 1998, 2000; Thorncroft and Jones 2000; Agustí-Panareda et al. 2004, 2005). The reintensification of a transitioning TC in response to an approaching upper-tropospheric disturbance is described by DiMego and Bosart (1982a,b) and others as similar to type-B EC development (Petterssen and Smebye 1971), whereby differential cyclonic vorticity advection produces surface development along a weak lower-tropospheric baroclinic zone.

A key aspect of the reintensification stage is that a TC may modify its environment such that thermodynamic and dynamic conditions are more favorable for extratropical cyclogenesis. For example, the reduced static stability and preexisting lower-tropospheric PV anomaly accompanying a recurving TC can facilitate a stronger interaction between an approaching upper-tropospheric PV anomaly and the lower-tropospheric PV anomaly of the TC (e.g., Thorncroft and Jones 2000). The stronger interaction promotes stronger enhancement of each PV anomaly by the other and, therefore, more rapid baroclinic development (Hoskins et al. 1985). Additionally, the presence of a recurving TC may enhance surface cyclone development ahead of an approaching upper-tropospheric trough by providing a source of lower-tropospheric cyclonic relative vorticity that subsequently can be enhanced through stretching (e.g., DiMego and Bosart 1982a,b; Bosart and Lackmann 1995). Furthermore, diabatically

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[3] Jones et al. (2003) expanded the definition of the second stage of ET to include the scenario of further cyclone decay, and thus renamed the second stage the "extratropical stage".

enhanced downstream upper-tropospheric ridge amplification accompanying a recurving TC can enhance quasigeostrophic forcing for ascent over the recurving TC by enhancing differential cyclonic vorticity advection and attendant upper-tropospheric divergence.

Finally, a downstream jet streak (e.g., Uccellini and Kocin 1987) induced by the recurving TC can promote extratropical reintensification (e.g., Sinclair 1993, 2004; Harr et al. 2000; Klein et al. 2002; Atallah et al. 2007). A strengthening meridional upper-tropospheric geopotential height gradient [alternatively manifested as a meridional upper-tropospheric PV gradient or potential temperature gradient on the dynamic tropopause (DT)] associated with downstream ridge amplification can produce a jet streak poleward and slightly downstream of the TC, thus placing the TC beneath the ascending branch of the jet entrance secondary circulation. Such a configuration can lead to mutual intensification of the jet streak and TC as long as the TC remains coupled with the jet streak.

iii. Influence of TC recurvature on the downstream extratropical flow pattern

A TC may modulate not only the nearby extratropical flow pattern as it recurves and undergoes ET, but the downstream extratropical flow pattern as well. For example, the lens of negative or low PV residing just below the tropopause associated with a TC may be deformed and advected downstream by the extratropical jet stream as a TC recurves. This process strengthens the upper-tropospheric meridional PV gradient associated with the extratropical jet stream (i.e., waveguide; e.g., Schwierz et al. 2004), resulting in a stronger and more zonally elongated jet stream.

The extratropical flow response to recurving WNP TC Jangmi (2008) provides an example of how a recurving TC may strengthen and zonally elongate the jet stream over the North Pacific. A study by Grams (2010) indicates that diabatic heating along the developing warm front of transitioning and recurving TC Jangmi generates upper-tropospheric outflow that strengthens and elongates the downstream jet stream. Using PV inversion (e.g., Davis and Emanuel 1991) to quantify the impact of the TC on the extratropical flow, Grams (2010) finds that if the TC is removed from the initial conditions of a model simulation, an upper-tropospheric trough develops instead of a zonal jet stream downstream of where the TC would be located. Thus, the upper-tropospheric outflow associated with TC Jangmi apparently acts to prevent upper-tropospheric trough formation downstream of the TC and leads to a strong, zonally elongated jet stream.

Another type of extratropical flow modification by a recurving TC is illustrated by the indirect disruption of the NH extratropical circulation caused by Hurricane Katrina (2005) (McTaggart-Cowan 2007b). Following the recurvature of Katrina, which undergoes ET but does not reintensify as an EC, the upper-tropospheric ridge associated with Katrina acts in conjunction with an extratropical upper-tropospheric trough to produce a corridor of upper-tropospheric southwesterlies over the U.S. The southwesterlies subsequently inject a pool of potentially warm air on the DT into the extratropics that persists for approximately two weeks and nearly circumnavigates the NH. The long-lived impact of Katrina on the NH flow pattern implies that individual episodes of recurving TCs can impact the extratropical circulation on intraseasonal time scales.

In addition to the aforementioned examples of how recurving TCs can modulate the extratropical flow, recurving TCs can perturb the extratropical jet stream/waveguide from its equatorward side such that a RWT is excited or amplified. RWTs occur frequently over the North Pacific in the cool season (e.g., Hakim 2003) and are typically excited by DT disturbances or upper-tropospheric troughs originating from Asia that may perturb the WNP jet stream/waveguide from its poleward side (e.g., Orlanski 2005). The high baroclinicity over the WNP established by the warm Kuroshio Current and relatively frequent incursions of cool Asiatic air masses favors frequent cyclogenesis events that can amplify the wave trains (e.g., Hoskins and Valdes 1990; Orlanski 2005). The wave trains then may disperse downstream along the North Pacific jet stream/waveguide, and can be amplified by further cyclogenesis events. Rossby wave amplification and dispersion is a known precursor to wintertime large-scale flow reconfigurations and attendant downstream high-impact weather events such as heavy precipitation events (e.g., Martius et al. 2008; Archambault et al. 2010). A caveat to considering RWT amplification and dispersion associated with recurving TCs is that the North Pacific baroclinicity and jet stream are considerably weaker in the summer and fall, when most episodes of TC recurvature occur, than in the winter.

Exploration of Rossby wave amplification and dispersion specifically associated with recurving WNP TCs has been conducted via individual case studies. An eddy kinetic energy budget analysis by Harr and Dea (2009) of four recurving and transitioning WNP TCs during 15 July–30 September 2005 reveals that two of the four TC recurvature events are associated with pronounced Rossby wave amplification and dispersion. Considerable variability in the Rossby wave response to the recurving WNP TCs is attributed mainly to differences in phasing between the TC and the extratropical flow. Harr and Dea (2009) further note that a transitioning TC may still induce Rossby wave amplification and dispersion without undergoing reintensification as an EC.

The interaction of a recurving TC with the extratropical flow has been examined via idealized modeling. Riemer et al. (2008) use a full-physics model with an outer domain and a two-way nested inner domain to show that a recurving TC interacting with a straight, extratropical jet stream initiates an upper-tropospheric ridge–trough couplet and jet streak just downstream of the TC (Fig. 1.5a). Extratropical cyclogenesis then occurs in the left-exit region of this downstream jet streak (Fig. 1.5b), which initiates a new ridge–trough couplet downstream, and so on (Figs. 1.5c,d). This sequence of events is consistent with downstream baroclinic development (e.g., Orlanski and Sheldon 1995). In the latter stages of the simulation (Figs. 1.5c,d), cyclonic and anticyclonic wave breaking (CWB and AWB; e.g., McIntyre and Palmer 1983) develops as the upper-tropospheric flow downstream of the recurving TC increasingly deforms.

Numerous case studies of ET (e.g., Hoskins and Berrisford 1988; Bosart and Lackmann 1995; Klein et al. 2000; Colle 2003; Agustí-Panareda et al. 2004; McTaggart-Cowan et al. 2006; Atallah et al. 2007) suggest that along with warm-air advection, diabatic heating associated with heavy precipitation accompanying the transitioning TC can act to amplify a downstream ridge and intensify a jet streak. The idealized modeling studies by Riemer et al. (2008) and Riemer and Jones (2010) demonstrate that the diabatically driven divergent outflow of the TC is essential to initial ridge amplification and jet streak intensification during TC recurvature. As a TC recurves, its diabatically driven divergent outflow builds and “locks in” a downstream upper-tropospheric ridge and jet streak. This interaction induces a downstream ridge–trough pattern. Once the ridge–trough pattern develops, the rotational TC outflow acts with the divergent TC outflow to amplify the downstream ridge by advecting low PV air into the ridge. The diabatic contribution to ridge amplification during TC recurvature implied by the role of divergent outflow in inducing an upper-tropospheric ridge is analogous to the diabatic ridge amplification that can accompany explosively deepening oceanic extratropical cyclones featuring widespread convection in their warm sector [e.g., the 12–14 March 1993 "Superstorm" along the eastern U.S. coast, Bosart et al. (1996); Bosart (1999)].

The importance of diabatically driven outflow in Rossby wave amplification accompanying TC recurvature as indicated in case studies and idealized modeling studies is corroborated by an analysis of ensemble forecast data for recurving WNP TCs Tokage (2004) and Nabi (2005) by Torn (2009). This study finds that downstream ridge amplification accompanying TC recurvature is highly sensitive to the amount of warm-frontal precipitation, lower-tropospheric moisture flux, and upper-tropospheric divergence associated with the TC. That the diabatically driven divergent outflow associated with a recurving TC promotes Rossby wave amplification is also consistent with the concept of negative vorticity advection by the divergent wind as a wave source for low-frequency RWTs emanating from the subtropics (Sardeshmukh and Hoskins 1988). The apparent importance of diabatic heating in Rossby wave amplification during TC recurvature is indicative of a predictability problem because model convective parameterization schemes are notoriously poor at capturing the bulk upscale effect of diabatic heating associated with convection (e.g., Dickinson et al. 1997; Bosart 1999; Langland et al. 2002; Zhang et al. 2003).

The sensitivity of the Rossby wave response to different environmental conditions and TC characteristics during TC recurvature has been tested in idealized model simulations (Riemer et al. 2008; Riemer and Jones 2010). Compared to TCs that recurve into a relatively weak jet stream, TCs that recurve into a relatively strong jet stream are found to be associated with longer-wavelength Rossby waves that are characterized by more distinct CWB and feature a faster eastward group velocity (Riemer et al. 2008). This finding is consistent with findings that Rossby wave dispersion during recurving WNP TC episodes may be favored when North Pacific lower-tropospheric baroclinicity is relatively strong (e.g., Reynolds et al. 2009).

Riemer et al. (2008) also find that switching off the parameterizations representing moist processes in the outer domain of the simulation results in a lower-amplitude RWT than when the parameterizations representing moist processes are left on. Riemer et al. (2008) attribute the lower-amplitude RWT in the absence of parameterized moist processes in the extratropics to the weaker extratropical cyclogenesis that occurs downstream of the recurving TC in the absence of parameterized diabatic heating. Further, Riemer et al. (2008) find that the initialization of the model simulation with a weaker and somewhat smaller TC than in the original experiment produces a lower-amplitude RWT, which is apparently the result of smaller and weaker upper-tropospheric TC divergent outflow impinging upon the jet stream. Of note is that the wavelength and group velocity of the RWT associated with the recurving TC in the simulations of Riemer et al. (2008) are not affected by either parameterized moist processes in the extratropics or by the size and intensity of the recurving TC.

A modeling study of six cases of Atlantic TCs that undergo recurvature and ET (Davis et al. 2008) indicates that in the presence of strong vertical wind shear, larger TCs are associated with greater vertical mass flux, and by inference, more precipitation, upper-tropospheric divergence, and diabatic heating, than smaller TCs. Davis et al. (2008) hypothesize that larger TCs are more likely to stay coherent as they interact with the jet stream, which enables them to induce higher-amplitude ridges along the extratropical jet stream than smaller TCs with weaker divergent circulations.

In an idealized experiment by Riemer and Jones (2010) in which a TC recurves in the presence of a preexisting extratropical RWT, the evolution of the extratropical flow pattern downstream of the recurving TC is found to be sensitive to the location of the TC relative to the RWT. Specifically, whether the recurving TC acts to enhance or dampen a RWT is found to be highly sensitive to the position of the TC relative to the preexisting trough–ridge pattern along the jet stream (i.e., to the extent of phasing between the TC and trough–ridge pattern). In this study by Riemer et al. (2008), TC recurvature into the leading edge of the RWT is identified as the optimal scenario in which to produce a high-amplitude, long-lived perturbation to the downstream extratropical flow pattern.

iv. Predictability issues associated with TC recurvature

Predicting the behavior of recurving TCs and their downstream impact has long been considered problematic for both human forecasters and numerical models (e.g., Riehl and Shafer 1944; Burroughs and Brand 1973, their Table 1; Chan et al. 1980; Cheng-Lan and Sadler 1983; Dobos and Elsberry 1993; Holland and Wang 1995; Jones et al. 2003). Several studies have found that human and model forecasts of TC track are less skillful for TCs that undergo recurvature than for those that do not (e.g., Riehl and Shafer 1944; Burroughs and Brand 1973; Chan et al. 1980; Dobos and Elsberry 1993). Since the presence of a recurving and transitioning TC in the extratropics can modulate the evolution of the downstream extratropical flow pattern, poor forecasts of TC recurvature and ET may lead to significant reductions in both regional and hemispheric predictability

Due to the complex interaction between recurving TCs undergoing ET and the extratropical flow, it has been hypothesized that the intrinsic predictability of recurving TCs undergoing ET is lower than that of pure TCs or ECs (e.g., Jones et al. 2003). Results of a study by Evans et al. (2006) help support this hypothesis. In their study, they find that model forecasts of cyclone structure verifying during ET tend to be poor relative to forecasts verifying before or after ET. Consistent with the finding of Evans et al. (2006), a study by Torn and Hakim (2009) using an ensemble-based analysis of recurving WNP TCs Tokage (2004) and Nabi (2005) shows high sensitivity of ET to initial condition error associated with the phasing between a recurving TC and an upstream trough.

Episodes of TC recurvature may coincide with skill reductions in model forecasts of the extratropical flow pattern over an entire ocean domain. Jones et al. (2003) note that the skill of 48-h, 72-h, 96-h, and 120-h forecasts of midtropospheric geopotential height for the North Pacific is substantially reduced during three episodes of WNP TC recurvature and ET in August 1996 (Figs. 1.6a–d, respectively). They conclude that the reduced forecast skill is related to model failure to capture the phasing between the TCs and upstream extratropical disturbances. The forecast errors associated with phasing translate to large forecast errors in the TC tracks, which lead to large forecast errors in the placement of synoptic features downstream of the TCs.

Forecast uncertainty associated with a recurving and transitioning TC may grow and spread downstream within the extratropical jet stream. This process can be visualized using a time–longitude plot of the standard deviation of the ensemble member forecasts of a given variable, often 500-hPa geopotential height (Harr et al. 2008; Anwender et al. 2008). Figure 1.7 shows this type of display generated from the ECMWF EPS initialized just prior to the recurvature of TC Tokage in October 2004. The term “plume of uncertainty” is used to describe an increase in magnitude and extent of forecast standard deviation emanating from a point on a time–longitude plot. A plume of uncertainty originating from the time and position of the recurvature of TC Tokage (Fig. 1.7) is apparently reinforced at the time and position of the completion of the ET of TC Tokage. Therefore, Fig. 1.7 indicates that both the recurvature and completion of ET of TC Tokage are associated with increased forecast uncertainty. By examining similar time–longitude plume diagrams for several recurving and transitioning TC episodes, Anwender et al. (2008) determine that forecast uncertainty often increases downstream of recurving and transitioning TCs over the WNP and North Atlantic basins.

Using a combination of principal component analysis and fuzzy clustering (Harr and Elsberry 1995), the range and likelihood of extratropical flow evolution scenarios associated with a given recurving TC episode based on EPS output can be synthesized (Harr et al. 2008; Anwender et al. 2008). This statistical technique groups EPS ensemble members based upon their contribution to the overall ensemble variability. Using this technique, Anwender et al. (2008) determine that forecast uncertainty in the extratropical flow evolution associated with recurving and transitioning TCs generally can be described by uncertainty in the phase, tilt, and amplitude of the extratropical wave pattern.

A sensitivity study of ensemble forecasts for the recurvature and ET of WNP TC Tokage by Anwender et al. (2010) suggests that the uncertainty in the amplitude of the downstream ridge associated with TC recurvature may arise from the high sensitivity of the ridge amplification to lower-tropospheric latent heat release along the developing warm front of the transitioning TC. The study by Anwender et al. (2010), as well as a study by Reynolds et al. (2009), also indicates that small initial condition error associated with TC recurvature and ET may spread downstream and grow with the RWT excited by the TC recurvature. Thus, TC recurvature episodes that excite or amplify a RWT may have a detrimental impact on predictability far downstream.

As discussed in the literature review above, individual case studies (e.g., Harr and Dea 2009; Grams 2010) and idealized modeling studies (i.e., Riemer et al. 2008; Riemer and Jones 2010) have shown that recurving TCs are often accompanied by downstream ridge amplification and jet stream intensification induced by the upper-tropospheric diabatically driven outflow of the TC that may lead to Rossby wave dispersion. However, an investigation of the characteristic downstream extratropical flow response to recurving TCs based on a large number of cases has not yet been performed. By employing a climatology and composite analysis to examine the extratropical flow response to recurving WNP TCs occurring over a 31-year period (1979–2009), this study will assess the synoptic–dynamic factors that modulate the observed downstream extratropical flow response to recurving WNP TCs.

An additional unique aspect of this work is that it will employ multiscale case studies to identify the key synoptic–dynamic processes (e.g., diabatically enhanced ridge amplification and jet streak intensification, CWB and AWB, and extratropical cyclogenesis) that link certain WNP TC recurvature episodes to high-impact weather events over North America. Furthermore, this work will complement previous statistical analyses of the predictability of recurving TCs (e.g., Harr et al. 2008; Anwender et al. 2008; Torn and Hakim 2009; Anwender et al. 2010) by combining conventional statistical assessments of predictability (e.g., EPS member spread) with analyses created from high-resolution gridded datasets in an examination of how synoptic–dynamic processes influence the predictability of the downstream extratropical flow response to recurving WNP TCs.

FIGURES

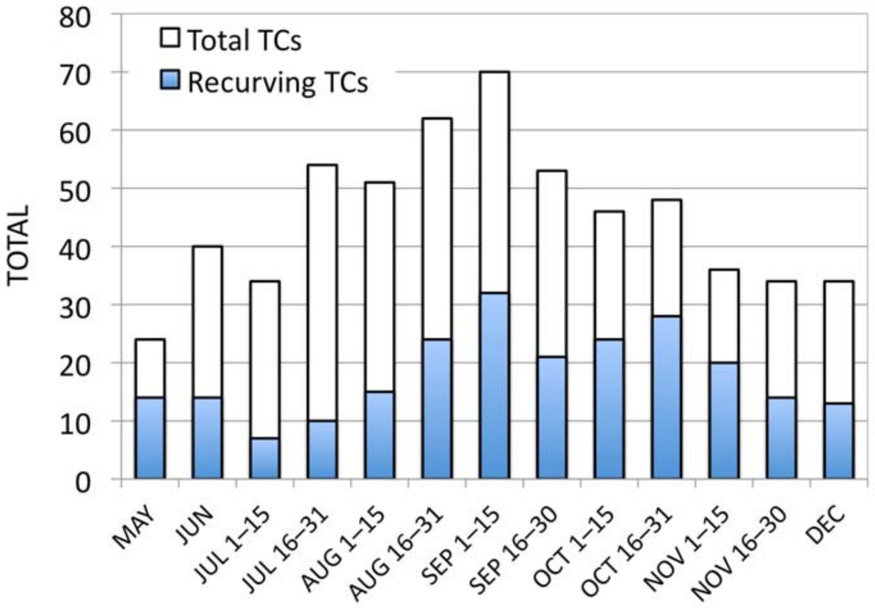

Fig. 1.1. A 1945–1969 climatology of the total WNP TCs and recurving WNP TCs [adapted from Table 2 of Burroughs and Brand (1973)]. TC recurvature is defined as a change in TC heading from generally westward to generally eastward.

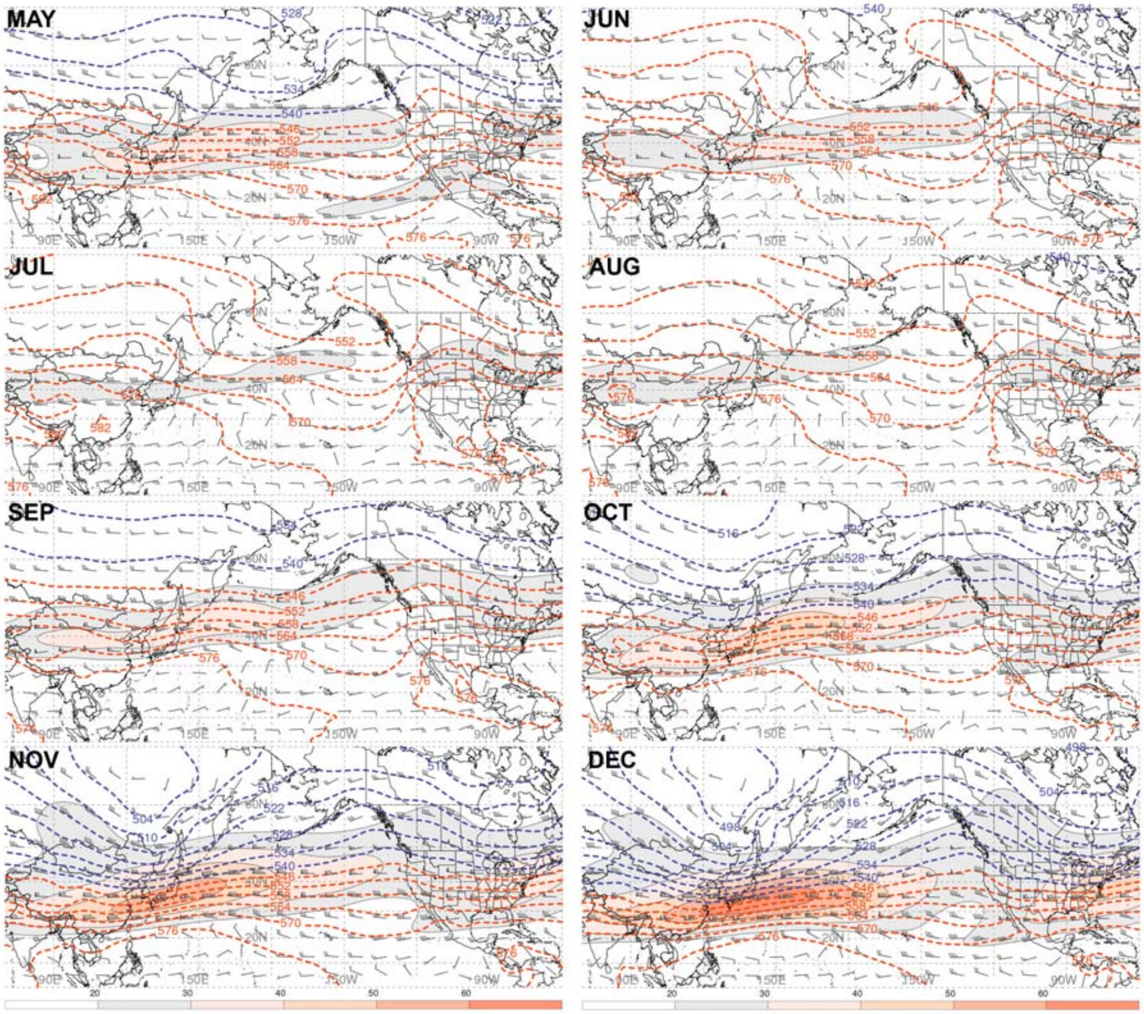

Fig. 1.2. The long-term (1979–2009) monthly mean 250-hPa wind speed (shaded in m s^{-1} according to the color bar), wind (barbs, kt) and 1000–500-hPa thickness (dashed, every 6 dam) over East Asia, the North Pacific, and North America for May–December.

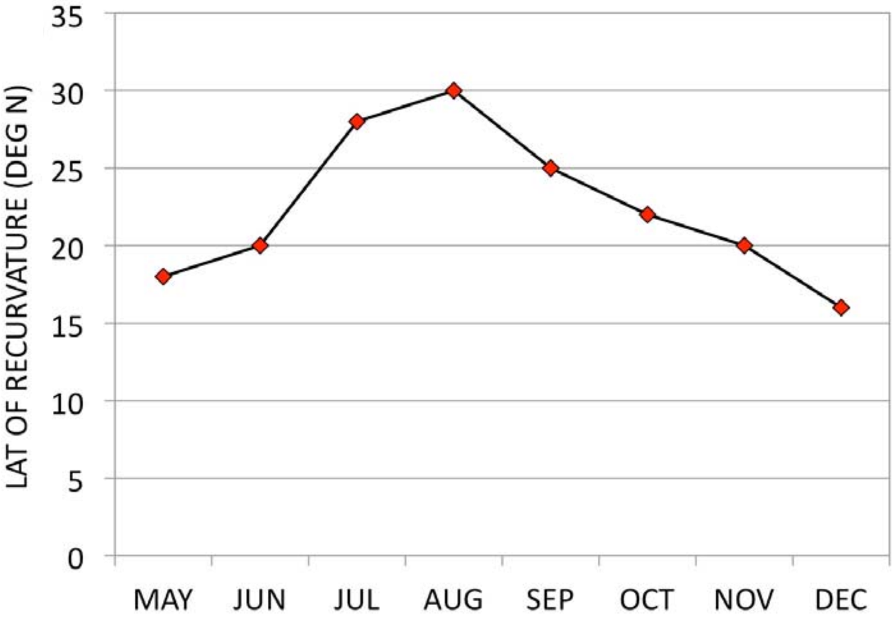

Fig. 1.3. A 1945–1969 climatology of the monthly mean latitude of WNP TC recurvature for May–December [adapted from Fig. 1a of Burroughs and Brand (1973)].

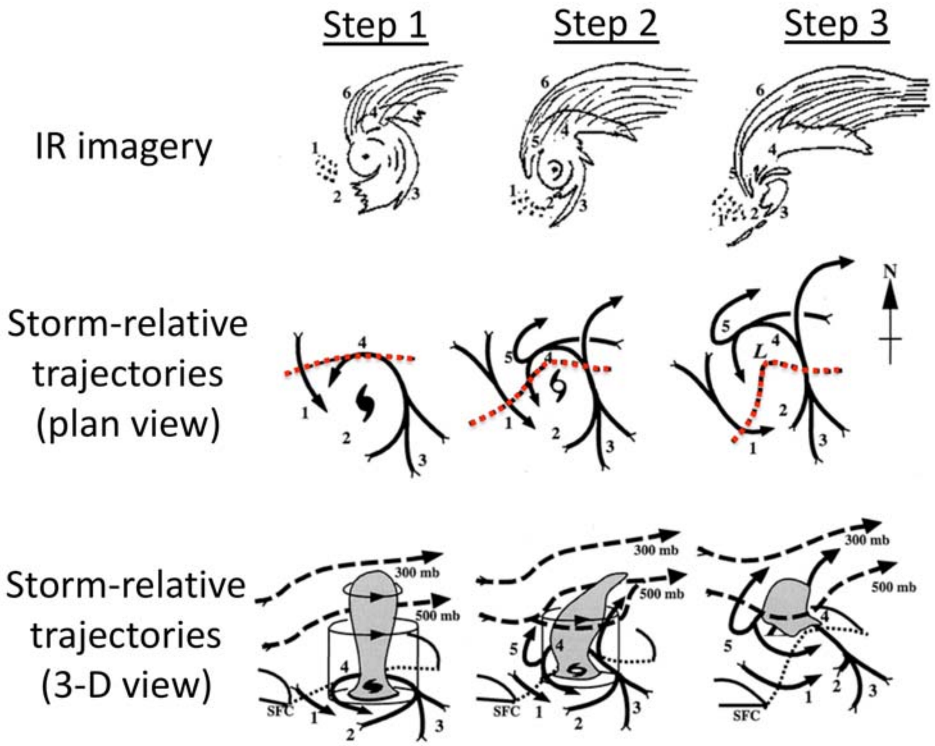

Fig. 1.4. Conceptual model of the transformation stage of ET. Dashed lines represent baroclinic zones, regions of gray shading denote the warm core of the cyclone, dashed arrows indicate polar jet streams, and solid arrows denote trajectories. The small numbers in each panel represent the following processes and features: (1) equatorward advection of cool, dry air; (2) decreased convection on the western side of the cyclone in step 1, which extends to the southern side of the cyclone in steps 2 and 3; (3) poleward advection of warm, moist air that maintains convection on the eastern side of the cyclone; (4) lower-tropospheric ascent of warm, moist air over the baroclinic zone; (5) midtropospheric ascent on the western side of the cyclone in steps 2 and 3; and (6) the cirrus shield. [Adapted from Fig. 5 of Klein et al (2000).]

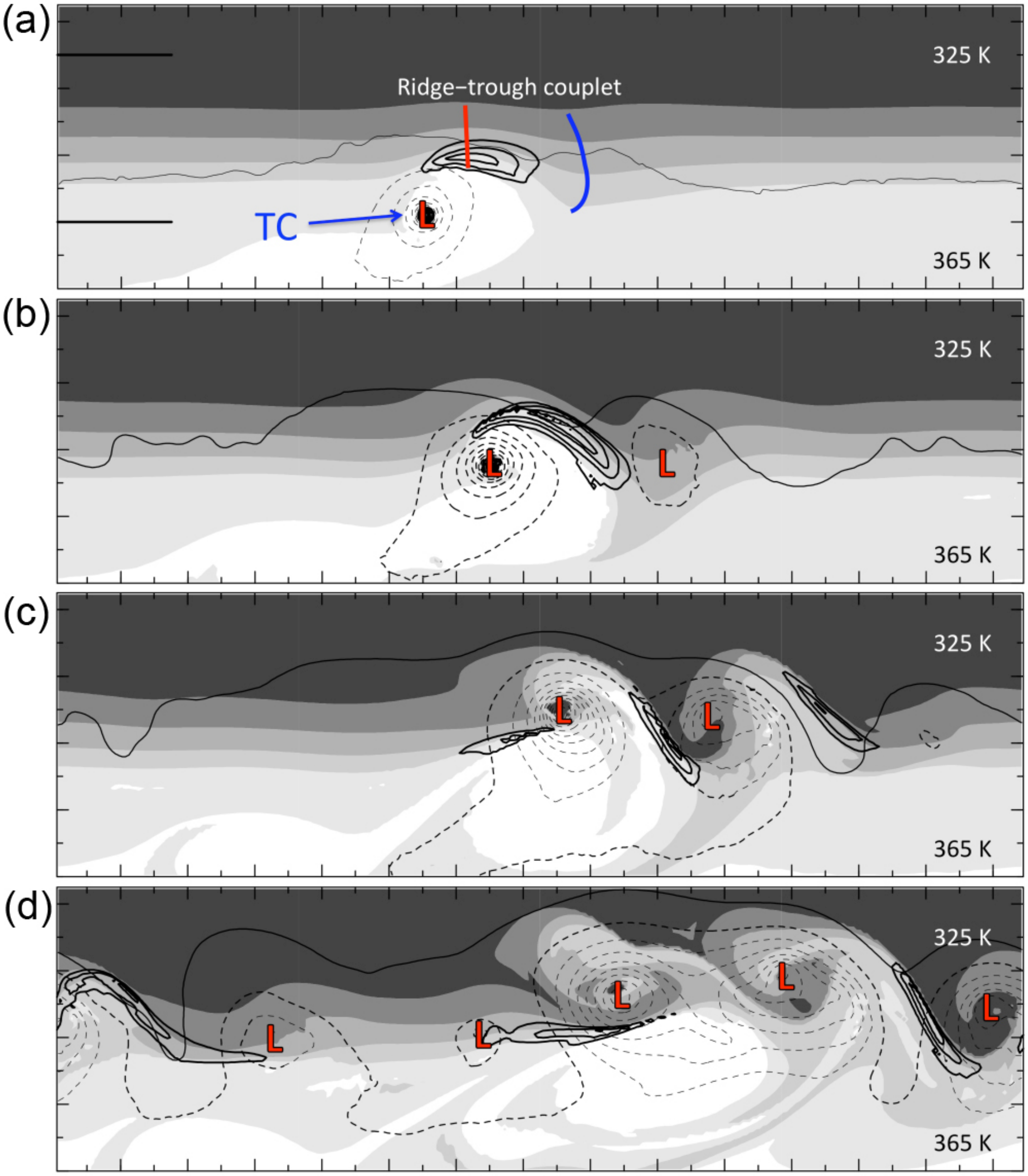

FIG. 1.5. A full-physics numerical model simulation showing the interaction of a recurving TC with an idealized straight jet at (a) 120 h, (b) 156 h, (c) 192 h, and (d) 240 h. Plots show the DT (i.e., 2-PVU surface) potential temperature (shaded every 10 K) and wind speed (thick contours; every 10 m s^{-1} starting at 45 m s^{-1}), and SLP (thin contours, every 5 hPa, values equal to or less than 995 hPa are dashed). The domain extends from 14°N to 65°N, with the jet stream axis at the start of the simulation located at 42°N. The outer model domain is 17 280 km by 8 460 km, and the two-way nested inner domain is 1 200 × 1 200 km. The distance between tick marks is 600 km. [Adapted from Fig. 2 of Riemer et al. (2008).]

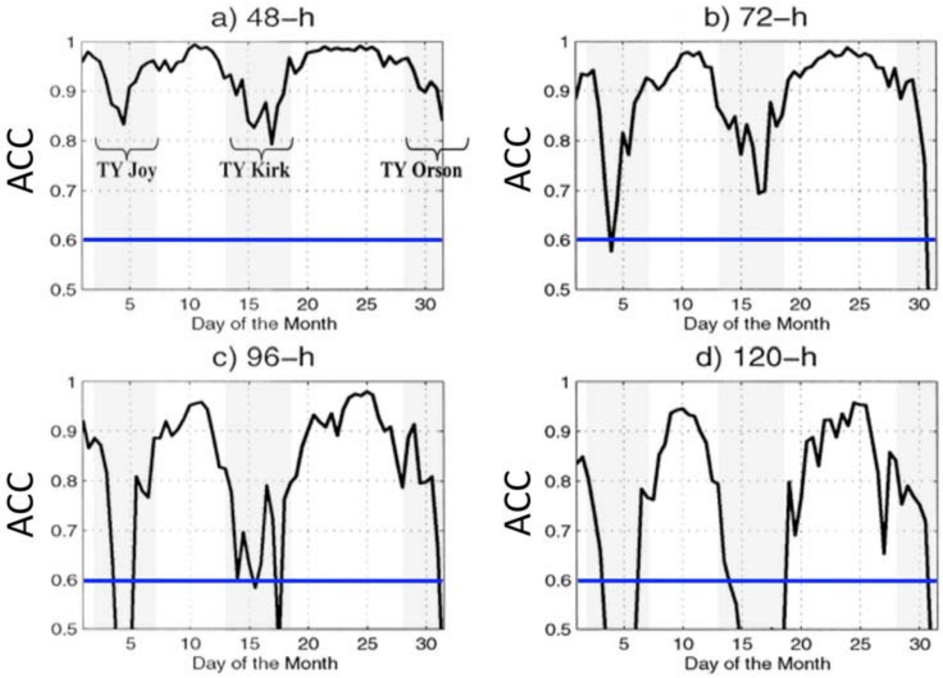

Fig. ic 500-hPa geopotential height forecasts in August 1996. Panels (a) through (d) show 48-h, 72-h, 96-h, and 120-h forecast ACCs, respectively. Three periods of WNP TC recurvature and ET are denoted in (a). Horizontal blue lines delineate the ACC threshold (0.6) above which a forecast is considered skillful. [Adapted from Fig. 8 of Jones et al. (2003).]

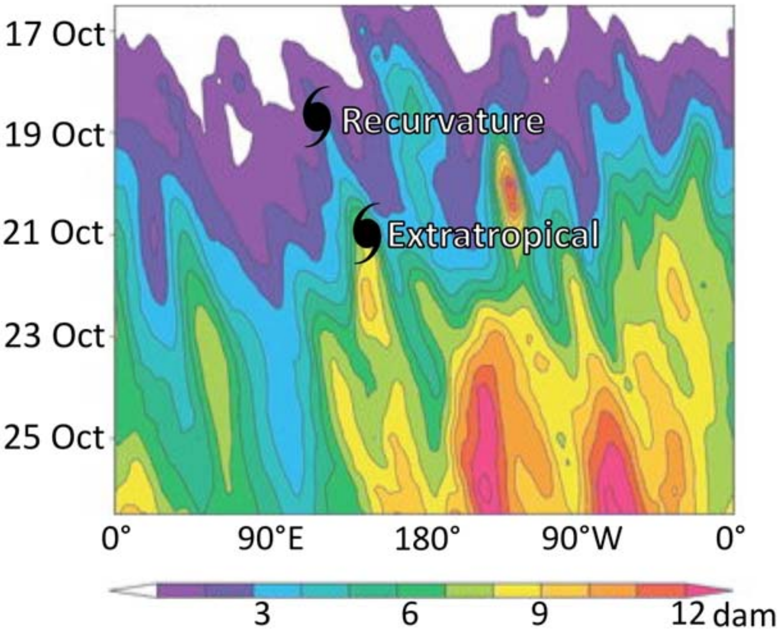

FIG. 1.7. Time–longitude plot showing the 40°–50°N averaged standard deviation of 500-hPa geopotential height forecasts (shaded every 10 m according to the color bar) from the 51-member ECMWF ensemble prediction system initialized at 1200 UTC 16 October 2004. The two typhoon symbols indicate where and when Tokage recurves and becomes extratropical, respectively. [Adapted from Fig. 6e of Anwender et al. (2008).]