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ABSTRACT

A significant number of tropical cyclones move into the midlatitudes and transform into extratropical cyclones. This process is generally referred to as extratropical transition (ET). During ET a cyclone frequently produces intense rainfall and strong winds and has increased forward motion, so that such systems pose a serious threat to land and maritime activities. Changes in the structure of a system as it evolves from a tropical to an extratropical cyclone during ET necessitate changes in forecast strategies. In this paper a brief climatology of ET is given and the challenges associated with forecasting extratropical transition are described in terms of the forecast variables (track, intensity, surface winds, precipitation) and their impacts (flooding, bush fires, ocean response). The problems associated with the numerical prediction of ET are discussed. A comprehensive review of the current understanding of the processes involved in ET is presented. Classifications of extratropical transition are described and potential vorticity thinking is presented as an aid to understanding ET. Further sections discuss the interaction between a tropical cyclone and the midlatitude environment, the role of latent heat release, convection and the underlying surface in ET, the structural changes due to frontogenesis, the mechanisms responsible for precipitation, and the energy budget during ET. Finally, a summary of the future directions for research into ET is given.

1. Introduction

The threat to life and property from tropical cyclones is well recognized. Even in countries that are not directly affected by these storms, considerable media coverage is given to significant tropical cyclone events. Often, the level of interest diminishes when the tropical cyclone moves to higher latitudes and begins to lose its tropical cyclone characteristics such that official tropical cyclone warnings are no longer posted or the storm is thought to be no longer a threat to life or property. However, a decaying tropical cyclone often evolves into a fast-moving and occasionally rapidly developing extratropical cyclone that produces intense rainfall, very large waves, and even hurricane-force winds. This extratropical transition (ET) of a decaying tropical cyclone may result in a storm that continues to pose a serious threat to land and maritime activities by extending tropical cyclone–like conditions over a larger area and to latitudes that do not typically experience such events—essentially, bringing the strong winds typical of major winter storms to midlatitude locations during a summer or autumn season. Extratropical transition poses a significant challenge to the forecaster. There is a high degree of uncertainty associated with predicting the timing

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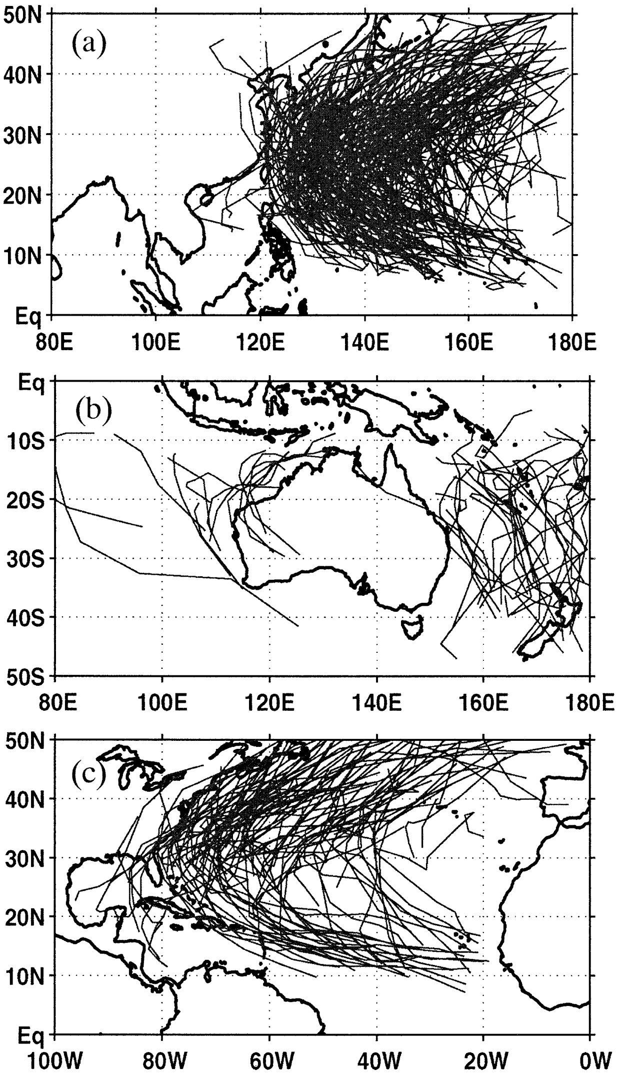

FIG. 1. Tracks of all tropical cyclones that underwent extratropical transition during 1970–99. (a) Western North Pacific. Tracks of tropical cyclones defined to be extratropical in JTWC best-track data. (b) Southwest Pacific [data as in (a)] and southeast Indian Ocean [tracks of tropical cyclones that accelerated toward the southeast under the influence of a midlatitude frontal system and maintained gales into midlatitudes, the so-called captured cyclones in Foley and Hanstrum (1994); best-track data taken from http://www.australiasevereweather.com/cyclones/history.htm]. (c) North Atlantic. Tracks of tropical cyclones defined to be extratropical in National Hurricane Center best-track data.

of the increased translation speed, the potential for rapid intensification, and the severity of the weather associated with ET, especially given the fact that the quality of numerical forecasts is not yet reliable for ET.

a. A brief climatology

Extratropical transition of a tropical cyclone occurs in nearly every ocean basin that experiences tropical cyclones (Fig. 1), with the number of ET events following a distribution in time similar to that of the total number of tropical cyclone occurrences (Fig. 2). The largest number of ET events occur in the western North Pacific (Fig. 2b) while the North Atlantic basin contains the largest percentage of tropical cyclones that undergo ET (Fig. 2a), with 45% of all tropical cyclones undergoing ET in the 30-yr period shown. In the eastern North Pacific the synoptic conditions associated with the presence of a strong subtropical ridge are not conducive to ET.

In the southwest Pacific, ET has a significant impact on Australia and New Zealand, being triggered by the approach of a midlatitude trough from the west (Sinclair 2002). However, because of the time frame shown, and due to the lack of a comprehensive definition of ET (as discussed later in this section), Fig. 1b underestimates the threat posed by ET to eastern Australia (J. Callaghan 2002, personal communication). Over the southeast Indian Ocean, relatively few tropical cyclones undergo ET (Figs. 1b and 2d). There, ET occurs most frequently when a large-amplitude cold front approaches within a distance of around 1700 km or less of a tropical cyclone. This is most likely to occur in the late summer or autumn. Such events result in significant impacts occurring every 20 yr or so (Foley and Hanstrum 1994).

Hart and Evans (2001) show that ET in the Atlantic occurs at lower latitudes in the early and late hurricane season and at higher latitudes during the peak of the season. The highest percentage of ET events occurs in September and October. The increased probability that Atlantic ET will occur during these months can be explained by comparing the geographical location of the areas that support tropical and extratropical development (see Hart and Evans 2001 for details of the calculation). In September and October the distance that a tropical cyclone must travel from the region that supports tropical development to the region that supports extratropical development is shorter than in other months, implying that a tropical cyclone is more likely to reach the region of extratropical development and thus more likely to undergo ET. Hart and Evans (2001) showed also that most of the storms that intensify after ET form in the deep Tropics and that a large number of these are Cape Verde cyclones.

Tropical cyclones that have undergone ET have been tracked across the Atlantic or the Pacific (e.g., Thorncroft and Jones 2000). Such systems may reintensify many days after ET and bring strong winds and heavy rain to the eastern side of the ocean basin [e.g., Hurricane Floyd in 1993 (Rabier et al. 1996); Hurricane Lili in 1996 (Browning et al. 1998)].

More detailed climatologies of ET are available for the west coast of Australia (Foley and Hanstrum 1994), the western North Pacific (Klein et al. 2000), the North Atlantic (Hart and Evans 2001), and the southwest Pacific (Sinclair 2002).

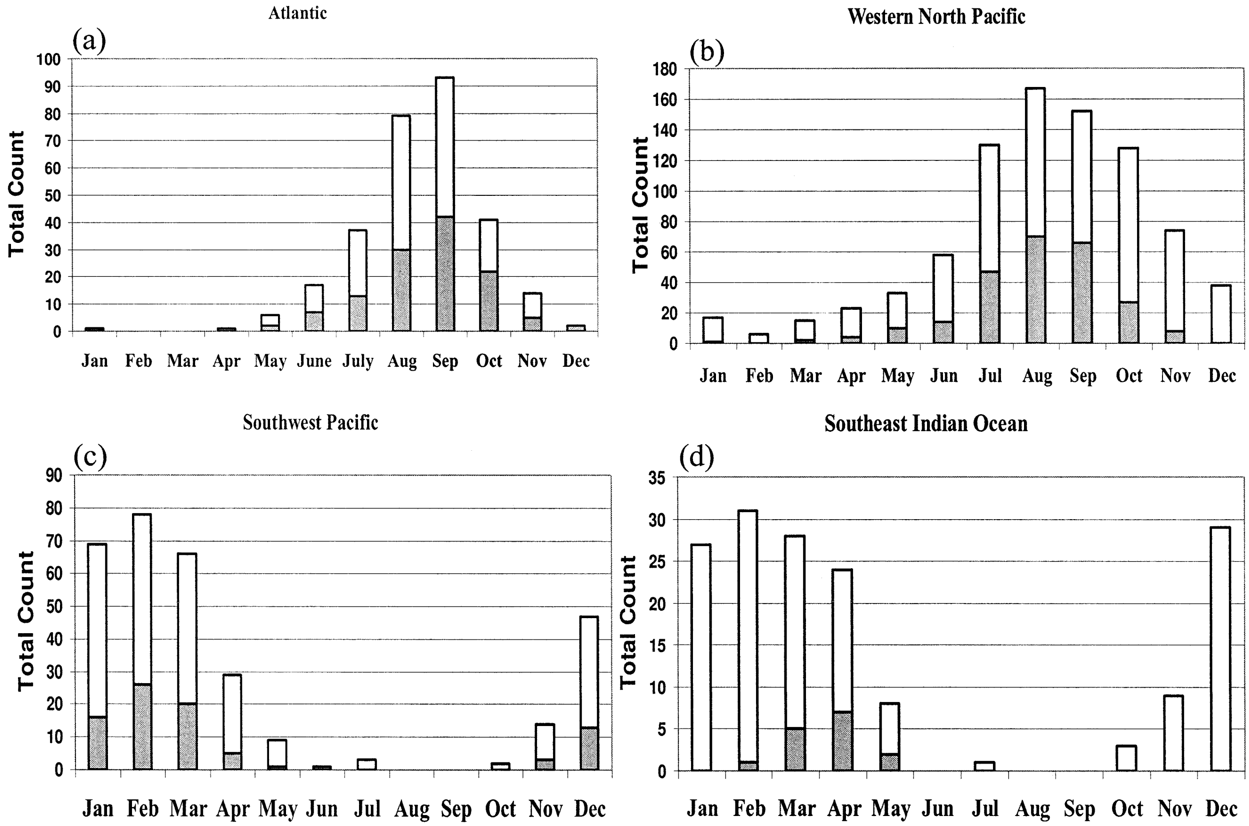

FIG. 2. Monthly total number of tropical cyclones (open bars) in each basin during 1970–99 and the number of tropical cyclones that underwent extratropical transition (shaded bars). Data as in Fig. 1.

b. Characteristics of extratropical transition

Extratropical transition is, as its name suggests, a gradual process in which a tropical cyclone loses tropical characteristics and becomes more extratropical in nature. As a tropical cyclone moves poleward it experiences changes in its environment (Schnadt et al. 1998). These changes may include increased baroclinity and vertical shear, meridional humidity gradients, decreased sea surface temperature (SST) or strong SST gradients (e.g., those associated with the Gulf Stream), and an increased Coriolis parameter. The tropical cyclone may come into proximity with an upper-level trough or a mature extratropical system. If the tropical cyclone makes landfall, it will experience increased surface drag, a reduction of surface fluxes of latent and sensible heat, and it may encounter orography.

When a tropical cyclone begins to interact with the midlatitude baroclinic environment, the characteristics of the cyclone change dramatically (Palmén 1958; Muramatsu 1985; Foley and Hanstrum 1994; Klein et al. 2000). In satellite imagery the inner core of the tropical cyclone loses its symmetric appearance and gradually takes on the appearance of an extratropical cyclone. The nearly axisymmetric wind and precipitation distributions that are concentrated about the circulation center of the tropical cyclone evolve to broad asymmetric distributions and expand greatly in area. Although the expanding cloud field associated with a poleward-moving tropical cyclone includes large amounts of high clouds due to the tropical cyclone outflow into the midlatitude westerlies, regions of significant precipitation are typically embedded in the large cloud shield.

Movement of a decaying tropical cyclone into the midlatitude westerlies results in an increased translation speed (Fig. 3), which contributes to the asymmetric distributions of severe weather elements. Over the ocean, high wind speeds and large translation speeds contribute to the generation of large ocean surface waves and swell (Bigio 1996; Bowyer 2000).

The southwest Pacific basin is unique in that tropical cyclones have an average *eastward* component of motion throughout most of their lives (Fig. 3). This is because they interact with the midlatitude westerlies early in their life cycle: these may extend to 15°S during the Southern Hemisphere tropical cyclone season. As a consequence, tropical cyclones start acquiring the asymmetries characteristic of the onset of ET (section 3a) around 20°S, closer to the equator than in any other ocean basin. On average, ET is complete by 30°S (Sinclair 2002). In contrast, Northern Hemisphere storms may preserve tropical characteristics as far north as 50°N.

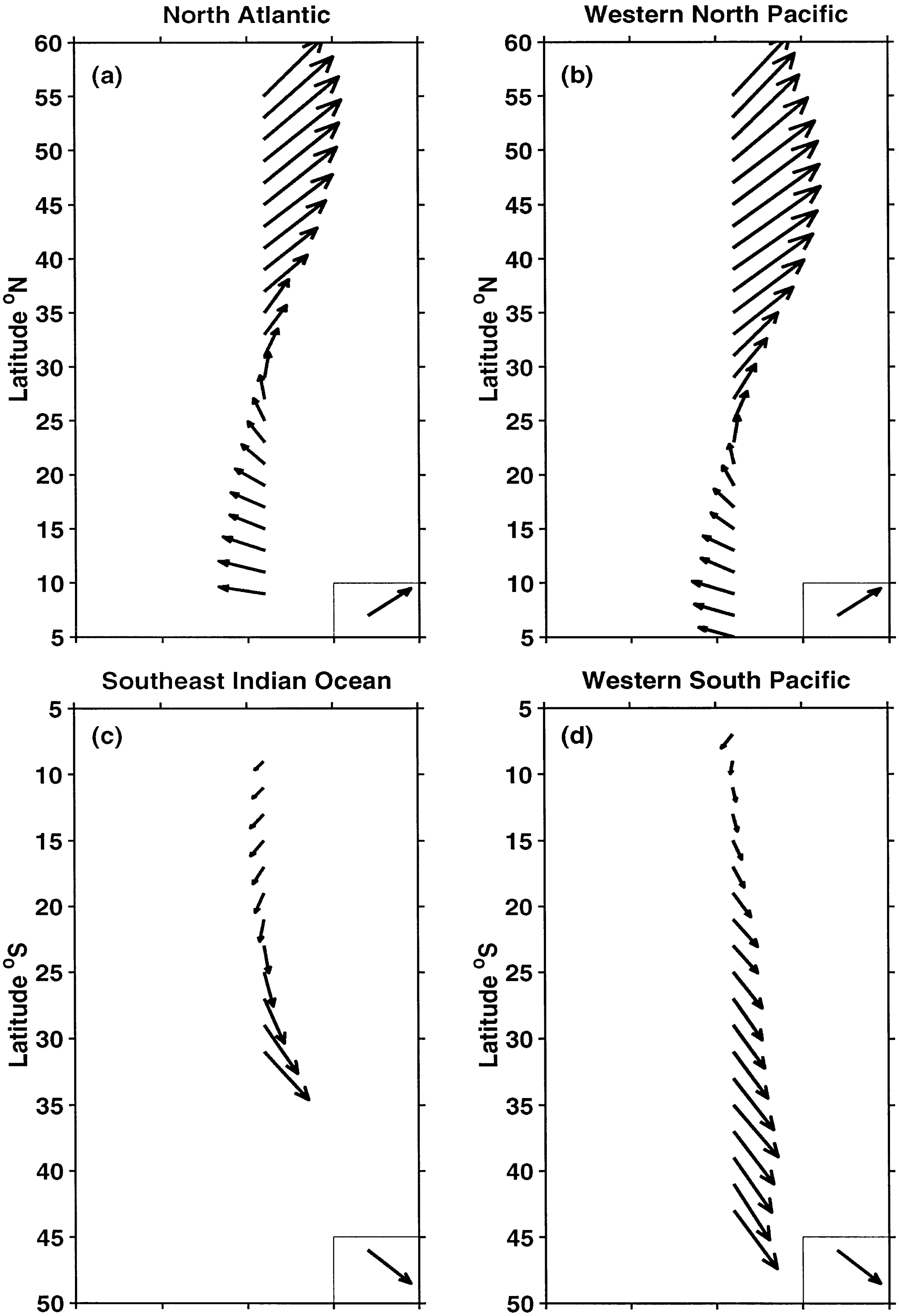

FIG. 3. Average motion of tropical cyclones vs latitude during 1970–99 from best-track data. The reference arrow has zonal and meridional components of 5 m s⁻¹.

c. Societal impacts

The societal impacts of ET can be substantial. In North America loss of life occurred because of severe flooding associated with the ET of Tropical Storm Agnes [1972 (Bosart and Dean 1991)]. In his study of the ET of Hurricane Hazel (1954), Palmén (1958) was the first to describe the rapid intensification and high amounts of precipitation that are associated with an ET over land. This case resulted in 83 deaths in the Toronto area of southern Ontario, Canada. In the northwest Pacific, severe flooding and landslides have occurred in association with ET. An example is the ET of Tropical Storm Janis (1995) over Korea, in which at least 45 people died and 22 000 people were left homeless (JTWC 1995). In one southwest Pacific ET event (Cyclone Bola) over 900 mm of rain fell over northern New Zealand (Sinclair 1993a). Another event brought winds gusting to 75 m s^{-1} to New Zealand's capital city, Wellington (Hill 1970), resulting in the loss of 51 lives when a ferry capsized. Extratropical transition has produced a number of weather-related disasters in eastern Australia, due to severe flooding, strong winds, and heavy seas [e.g., Cyclone Wanda in 1974 (Nicholls et al. 1998)]. In Western Australia there can be an increased fire hazard associated with ET. Tropical systems that reintensify after ET in the North Atlantic constitute a hazard for Canada [e.g., Hurricane Earl in 1998 (McTaggart-Cowan et al. 2001)] and for northwest Europe. The extratropical system that developed from Hurricane Lili (1996) was responsible for seven deaths and substantial economic losses in Europe.

d. Operational definition of extratropical transition

There is no commonly accepted definition of ET. A variety of factors are assessed by different forecast centers to decide whether or not a tropical cyclone is undergoing ET. In satellite imagery, indications of ET are a decrease in deep convection, the disappearance of the high-cloud canopy of a tropical cyclone, the exposure of the low-level circulation center, and the appearance of a comma-shaped cloud pattern or frontal structure (Merrill 1993). Further indications include an increase in the radius of gale force winds, asymmetries in the wind and precipitation fields, and a decrease in sea surface temperature beneath the tropical cyclone. Operational forecasting centers may continue to use the name assigned to the tropical cyclone during ET so that the general public does not underestimate the hazards associated with an ET event (e.g., in Canada an ET system is referred to as ''post tropical cyclone'').

e. Aims of this paper

Extratropical transition begins with a tropical cyclone and ends with an extratropical cyclone. Thus the challenges involved in forecasting and understanding ET contain elements of both tropical and extratropical meteorology. Forecasters responsible for tropical cyclone warnings over each basin in which ET occurs have perceived a need for increased understanding of the evolution of a tropical cyclone to an extratropical cyclone (WMO 1998, 2000). The purpose of this paper is not to provide a comprehensive review of either tropical or extratropical cyclones, but rather to summarize the current understanding of ET and to identify outstanding research and forecast issues. In section 2, forecasting ET is examined in terms of the challenges associated with predicting the impacts of severe weather. The current understanding of the characteristic life cycle of an ET event is described in section 3 with a summary of outstanding questions and future directions in section 4.

2. Forecast challenges and impacts

Forecasters in each regional center responsible for producing warnings and advisories during an ET event are faced with very similar problems in accurately predicting the behavior of the rapidly changing circulation during ET. The forecaster must predict not only the track and intensity of the system, but also the strength and distribution of surface winds and precipitation as well as the wave height if ET occurs over the ocean. Warnings must be issued for impacts such as flooding or bush fires. The majority of ET events occur over the ocean and satellite diagnostic techniques are therefore an important forecast tool. The prediction of ET is a major challenge for a numerical weather prediction model. An ET event can substantially reduce the skill of the medium-range forecasts downstream of the tropical cyclone and, thus, can have an impact on Europe and western North America. A further challenge for the forecaster is communicating with emergency management personnel and the public, particularly since there is much less public awareness of the hazards associated with ET than of those due to a tropical cyclone. In this section we discuss the forecast challenges and impacts associated with an ET event.

a. Track

A major challenge in forecasting the track of a tropical cyclone undergoing ET is the increase in its forward speed. As a tropical cyclone moves into the midlatitude westerlies, the contribution to the tropical cyclone motion from the environmental flow increases and the tropical cyclone can accelerate from a forward speed of 5 m s^{-1} in the Tropics to more than 20 m s^{-1} in the midlatitudes (Fig. 3). This acceleration compounds the forecast difficulty associated with predicting the location and duration of hazardous weather events such as large amounts of precipitation, high wind speed, generation of high surf, and large ocean wave heights and swell. The rapid change in translation speed decreases the warning time for small fishing and recreational vessels that frequent the marine areas in summer and autumn. If the timing of the increase in translation speed is misjudged, track errors of hundreds of kilometers can occur.

b. Intensity

The movement of a tropical cyclone into the midlatitudes is usually accompanied by a decrease in intensity as defined by an increase in central mean sea level pressure and a decrease in maximum surface wind speed. This decrease in intensity may result from reduced surface fluxes if the cyclone moves over land or over cooler water, from loss of symmetry following the intrusion of dry air into the circulation, or from the presence of environmental shear. However, the interaction with an extratropical system during ET may result in rapid reintensification as an extratropical system. This presents a major forecast challenge that is complicated by the fact that, if a forecaster is not aware of the possible interaction with an extratropical system, the satellite imagery would suggest that the tropical cyclone is weakening (Foley and Hanstrum 1994; Miller and Lander 1997). Rapid reintensification during an ET event can occur both over land [the ''Sydney'' cyclone in 1950 (Bureau of Meteorology 1950); Cyclone Audrey in 1964 (Bureau of Meteorology 1966); Hurricane Hazel in 1954 (Hughes et al. 1955; Matano 1958; Palmén 1958)] and over water [e.g., Cyclone Herbie in 1988 (Foley and Hanstrum 1994); Hurricane Emily in 1987 (Case and Gerrish 1988); Hurricane Iris in 1995 (Thorncroft and Jones 2000); Hurricane Earl in 1999 (McTaggart-Cowan et al. 2001); Hurricane Irene in 1999 (Prater and Evans 2002; Agusti-Panareda et al. 2003, manuscript submitted to *Quart. J. Roy. Meteor. Soc.*, hereafter APQJR)]. Alternatively, a tropical cyclone may decay significantly on entering midlatitudes, but the remnants of the tropical cyclone may interact with an extratropical system many days later. This kind of development, which can bring severe weather to western Europe and the North Pacific coast of North America, is often poorly forecast by numerical weather prediction models [e.g., Hurricane Floyd in 1993 (Rabier et al. 1996)].

Traditional measures used to assess tropical cyclone intensity and intensity change may not be appropriate during ET. Because of the strong asymmetries in the cloud field, the Dvorak (1975, 1984) classification of tropical cyclone intensity underestimates the maximum winds during the initiation of an ET event (Miller and Lander 1997). When reintensification occurs as an extratropical cyclone, the minimum sea level pressure attained by the extratropical system may be lower than that when the circulation was a tropical cyclone. However, the maximum winds in the tropical cyclone would be larger than those in the extratropical cyclone. This is a reflection of the fact that the extratropical system is moving to a region of lower background pressure (Sinclair 1997).

c. Surface winds

Knowledge of the nature of the low-level winds during ET is necessary to provide accurate warnings. The wind field in a tropical cyclone generally contains a weak wavenumber-1 asymmetry associated with the addition of the tropical cyclone motion to a symmetric wind field. Due to this asymmetry the strongest winds occur to the left (right) of the track in the Southern (Northern) Hemisphere (Powell 1982). During ET the system undergoing transition accelerates and the wind field typically decreases in strength resulting in a much stronger asymmetry than that observed in a tropical cyclone (Merrill 1993). In addition, the area covered by strong winds increases (Merrill 1993).

Over the ocean, considerably stronger winds have been observed by instruments situated higher up in the boundary layer (e.g., on offshore oil drilling platforms) than near the surface (e.g., at 10 m). Canadian reconnaissance discovered a low-level jet with maximum winds up to 72 m s^{-1} during of the ET of Hurricane Michael in 2000 (Abraham et al. 2002). Thus, the strength of the surface winds or gusts may be locally higher under convective cells or over warm ocean eddies where enhanced convection occurs. Over land, a similar reduction in surface wind speed may occur in association with frictional influences and the development of a stable surface layer. However, in the presence of deep dry convective mixing the frictional reduction of wind speed is much smaller. This is observed, for example, when an ET event occurs over central Australia where the mixed layer can extend to 4 km above the surface.

In view of the changes in the wind field during ET it is an important challenge for forecasters to predict how the structure of the wind field will change from that typical of a tropical cyclone, in which the strongest winds occur close to the center of the system, to the wind structure typical of a cyclone undergoing ET, in which the strongest winds occur over a much wider area and exhibit more significant asymmetries.

d. Precipitation and flooding

The precipitation associated with an ET event can be substantial and result in severe flooding. For example, northern Japan, Korea, and China are at risk of torrential rains from ET events, with flooding and landslides being more of a threat than strong winds [e.g., Typhoon Seth in 1994 (JTWC 1994); Tropical Storm Janis in 1995 (JTWC 1995)]. Natural disasters such as the ETs of Hurricane Floyd in 1999 (Atallah and Bosart 2003), Tropical Storm Agnes in 1972 (DiMego and Bosart 1982a,b), Hurricane Hazel in 1954 (Palmén 1958), and the 1938 Hurricane (Pierce 1939) were associated with interactions of decaying tropical cyclones with the baroclinic environment in the midlatitudes, resulting in extreme precipitation totals of 200–300 mm over a period as short as 18 h.

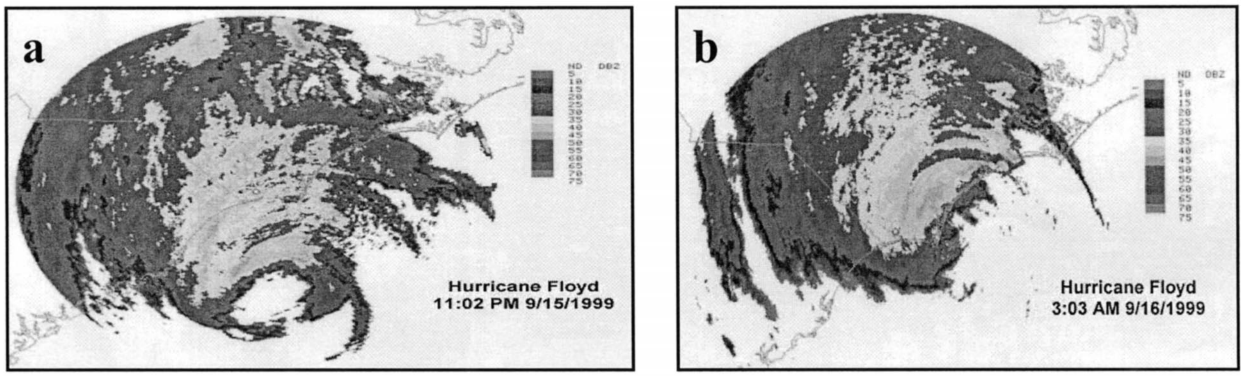

FIG. 4. Doppler radar reflectivity from the Wilmington, NC, National Weather Service Office at (a) 0300 and (b) 0700 UTC 16 Sep 1999. (Radar images are from http://nwsilm.wilmington.net/tropics/past_storms/1995_1999/1999/floyd/floyd.html.)

At the start of an ET event, heavy precipitation becomes embedded in the large cloud shield associated with the tropical cyclone outflow that extends poleward from the tropical cyclone center (Bosart and Dean 1991; Harr and Elsberry 2000; Kitabatake 2002). The heavy precipitation poleward of the tropical cyclone (Fig. 4) is not always anticipated since it often begins far from the tropical cyclone center. Due to the expansion of the area covered by clouds and precipitation when the tropical cyclone moves poleward, heavy precipitation can occur over land without the tropical cyclone center making landfall. If the heavy precipitation associated with the central region of the tropical cyclone then falls in the same region as the prestorm precipitation, the potential for flooding is increased.

In a tropical cyclone heavy precipitation typically occurs on both sides of the track (Frank 1977; Marks 1985; Rodgers et al. 1994), as seen for Cyclone Audrey (1964) directly after landfall in northern Australia (Fig. 5a). As ET proceeds, the distribution of heavy precipitation begins to resemble that of an extratropical cyclone with the heaviest precipitation to the right (left) of track in the Southern (Northern) Hemisphere (Fig. 5; Bergeron 1954; Palmén 1958; Carr and Bosart 1978; Bosart and Carr 1978; Bosart and Dean 1991; Foley and Hanstrum 1994). The change in the distribution of precipitation can be related to the large-scale forcing of ascent due to the interaction of the tropical cyclone with an upstream trough and/or with a low-level baroclinic zone (Bosart and Dean 1991; Harr and Elsberry 2000). The mechanisms responsible for the asymmetric precipitation are discussed in more detail in section 3h.

Orographic effects can enhance the precipitation during an ET event considerably. This is observed over Japan, the east coast of the United States, and over New Zealand. Orographic enhancement was a key factor in the phenomenal precipitation over New Zealand (more than 900 mm) from Cyclone Bola (1988) as warm, moist air with a horizontal wind speed of 25 m s^{-1} ascended local mountains (Sinclair 1993a). A simple orographic precipitation model was able to replicate this precipitation extremely well (Sinclair 1994). In the United States cold-air damming to the east of the Appalachians can lead to enhanced precipitation during an ET event (Bosart and Dean 1991; Atallah and Bosart 2003).

Extratropical transition poses an especially challenging quantitative precipitation forecasting (QPF) problem. Successful QPF requires an accurate prediction of the track, intensity, and structural changes of storms undergoing ET. The timing of the precipitation shift relative to the storm track described above is very sensitive to physical mechanisms that govern the ET process (e.g., the dynamical and thermodynamic structure of the upstream trough). Local and regional geography (e.g., coastlines and mountains) can also contribute significantly to the ultimate distribution of precipitation shields associated with storms undergoing ET. Current operational prediction models are often ''stressed'' when making QPF forecasts for storms undergoing ET. Likewise, simple rules of thumb that invoke precipitation duration as the dominant predictor for total storm rainfall may fail spectacularly for an ET event that may contain significant embedded mesoscale circulations in the presence of highly varied terrain.

e. Bush fires

The asymmetric nature of the wind and precipitation fields during ET leads to a bush fire hazard over Australia (Foley and Hanstrum 1994) as the strongest winds are associated with hot, dry air to the north. During the ET of Cyclone Alby in 1978 (Fig. 6) 114 000 ha of farm and forestland were destroyed by uncontrollable bush fires. Most of the fires were either already alight or had been deliberately lit on the day of the event as part of the traditional burning off of agricultural land in autumn prior to the onset of the winter rains. Timely warning could have prevented a lot of the fire damage that resulted. By comparison, 48 h of advance notice was given to fire authorities in March 1990 prior to the passage of Cyclone Vincent. This allowed existing fires

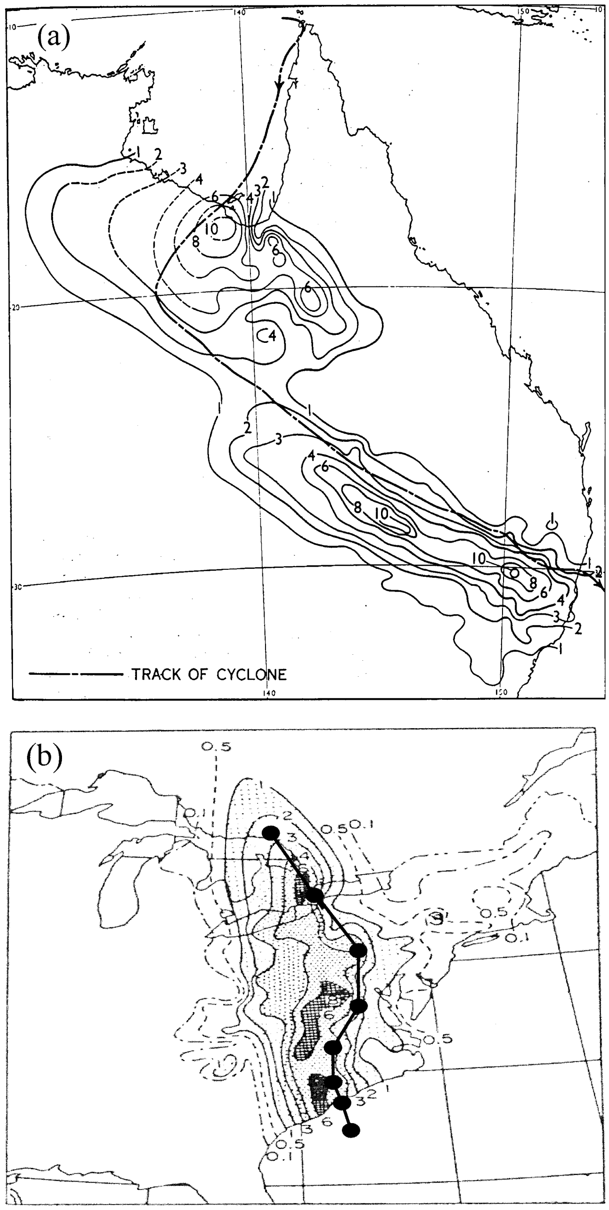

FIG. 5. Precipitation (in.) and cyclone track for the extratropical transition of (a) southwest Pacific Cyclone Audrey (1964) during the 72-h period ending 2300 UTC 14 Jan 1964 (taken from Bureau of Meteorology 1966) and (b) North Atlantic Hurricane Hazel (1954) during the 24-h period ending 0600 UTC 16 Oct 1954 [adapted from Palmén (1958)]. Solid circles mark the track of Hurricane Hazel in 3-h increments from 0900 UTC 15 Oct to 0600 UTC 16 Oct 1954.

to be suppressed and additional fire-fighting staff to be deployed during the period of maximum fire risk.

f. Ocean response

The occurrence of an ET event offshore can produce extremely large surface wave fields due to the continued high winds speeds and increased translation speed of the entire system. For slow-moving tropical cyclones the waves quickly move out ahead of the storm and advance as decaying swells. Nautical institutes worldwide teach mariners that one of the ''saving graces'' of tropical cyclones is the ground swell that advances well ahead of an approaching storm, thereby affording considerable advance warning. This is often not the case during ET when the faster-moving storm can arrive at the same time as the waves being generated, offering little or no advance warning. Wave heights 50%–100% higher than within a similar stationary system can result in a severe threat to any marine activity.

Figure 7 shows the wave heights associated with the extratropical transition of Atlantic Hurricane Danielle (1998). Significant wave heights measured by an offshore buoy increased by 9 m in 2 h (Fig. 7a). At the time at which Danielle was nearest the buoy (0800 UTC on 3 September), significant wave heights reached nearly 16 m with maximum waves exceeding 27 m. The extremely tight gradients at the leading edge of the wave field seen in Fig. 7b are typical for ET events. For example, peak waves of over 30 m occurred during the ET of Hurricane Luis (1995) and caused extensive damage to the luxury liner *Queen Elizabeth II* (Bigio 1996; Met Office 1996; Desjardins et al. 2000; Lalbeharry et al. 2000).

MacAfee and Bowyer (2000a,b) suggest that the high forward speeds of ET events can result in a resonance between the ocean waves being generated and the wind system generating them. In essence, the storm moves with the fetch, ''trapping'' the waves within the wind system, thereby allowing them to grow much larger than would be possible if the storm were stationary. The enhancement occurs only where the direction of storm motion is in the same sense as the wind, that is, on the right- (left-) hand side of the storm track in the Northern (Southern) Hemisphere. Storm–wave resonance theory has been generally understood for some time (e.g., Suthons 1945). However, its relevance to ET has been recognized only recently. Accordingly, the waves associated with a tropical cyclone undergoing ET can pose a greater threat than those associated with a stronger tropical cyclone within the Tropics.

g. Numerical prediction of extratropical transition

The primary guidance to an operational forecaster is often the numerical forecast model. The relatively small scale of the tropical cyclone and the complex physical processes that occur during the interactions between a tropical cyclone and the midlatitude environment present difficult numerical forecast problems. There are several limited-area numerical models that provide operational tropical cyclone track and intensity forecasts (e.g., Ueno 1989; DeMaria et al. 1992; Bender et al. 1993; Davidson and Weber 2000; Weber 2001). However, limited-area models that have adequate resolution to simulate the internal tropical cyclone features must rely on initial and boundary conditions from global

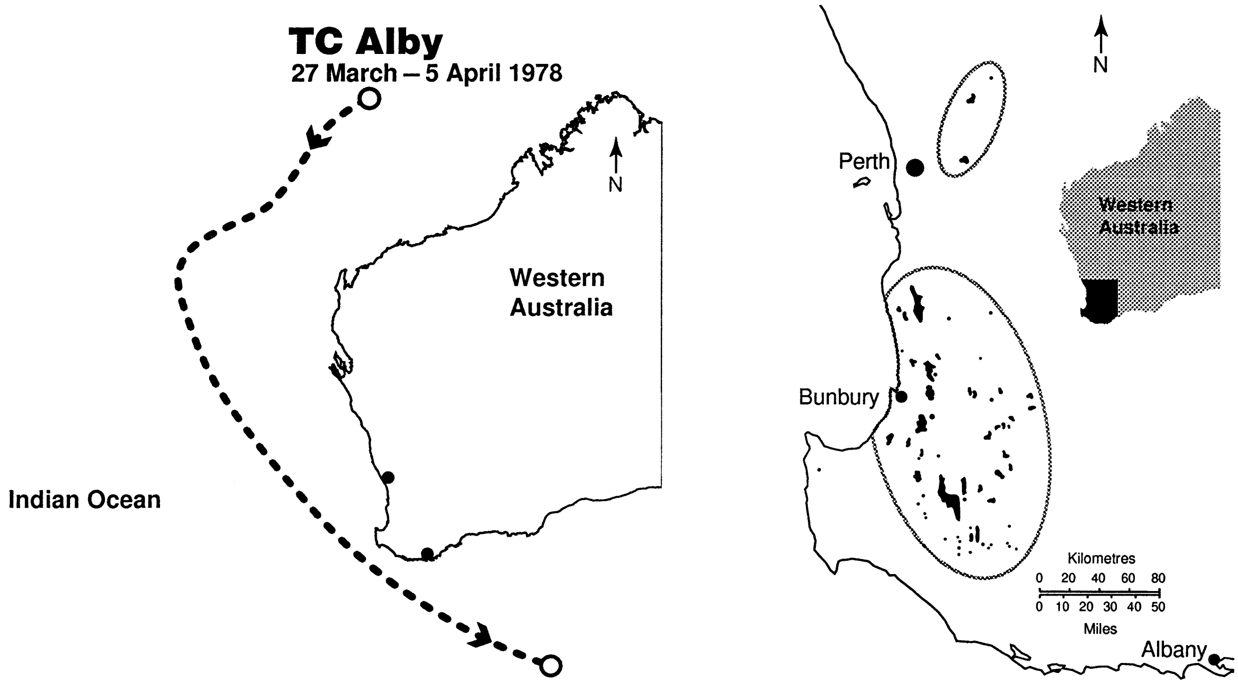

FIG. 6. Location of rural and forest areas burnt in bush fires associated with Cyclone Alby (1978). The dark shading shows the actual areas burnt; the ellipses indicate the regions affected by the fires. A total of 114 000 ha of forest and farming area in southwestern Australia was burnt.

models to represent the midlatitude circulation features into which the decaying tropical cyclone is moving. On the other hand, operational global models generally provide accurate guidance with respect to the midlatitude circulation patterns, but fail to capture the detailed tropical cyclone characteristics. Therefore, neither class of models adequately simulates interactions between a tropical cyclone and the midlatitude circulation.

The occurrence of an ET event often compromises the forecast skill across an entire ocean basin. During August 1996 (Fig. 8) the forecast skill of the Navy Operational Global Atmospheric Prediction System (NOGAPS) over the western North Pacific was reduced dramatically when Typhoon Joy, Typhoon Kirk, and Typhoon Orson recurved and underwent ET. For each case of ET, the decrease in anomaly correlation scores was related to the phasing in the forecast between the decaying tropical cyclone and the midlatitude circulation into which the tropical cyclone was moving. Errors associated with an incorrect phasing between the tropical cyclone and the midlatitude circulation often result in large location and intensity errors associated with the ET event, especially during reintensification of the decaying tropical cyclone as an extratropical cyclone. In particular, the large position errors (see discussion of Fig. 10 below) result in low anomaly correlation scores due to the misplacement of significant circulation features both associated with and downstream of the ET event. The magnitude of the decrease in skill seen in Fig. 8 became larger with the increase in forecast interval. Therefore, the numerical forecast accuracy for ET events may be far below the accuracy for tropical or midlatitude cyclones.

A characteristic problem associated with the numerical prediction of ET is that a large variability may occur in the representation of an ET event by different sets of forecasts that are initialized from sequential analyses. Forecasts of the central sea level pressure during the ET of Typhoon Bart (1999) give an example of this variability. The central pressure in the forecast from the global spectral model of the Japan Meteorological Agency (JMA) initialized at 0000 UTC 23 September 1999 (Fig. 9a) was slightly higher than the analyzed central pressure in the short range, but the deepening as an extratropical cyclone after 1200 UTC 24 September was too strong in the model forecast. The central pressure in the forecast initiated at 1200 UTC 23 September remained higher than in the analysis until 1200 UTC 26 September, when the model correctly forecast the minimum central pressure of the extratropical cyclone that formed from the ET of Typhoon Bart. In the next forecast sequence from 0000 UTC 24 September the model overdeveloped the extratropical cyclone and the minimum central pressure was reached 12 h too early. In ensuing forecasts, the central pressure was generally too deep during the intensification of the extratropical cyclone and too deep also during the weakening of the extratropical cyclone. Forecasts of the ET of Ty-

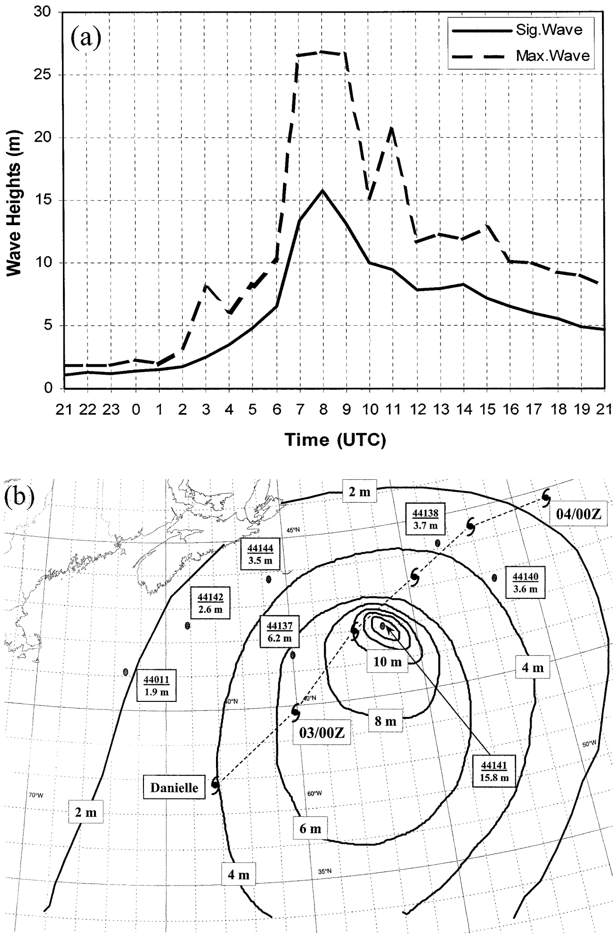

FIG. 7. (a) Significant and maximum wave heights reported from Canadian NOMAD buoy 44141 during the passage of Hurricane Danielle (1998). The largest waves were reported at the point when the storm was nearest the buoy indicating strong resonance between the storm and waves. Significant wave heights grew 9 m in only 2 h, indicating little advance warning of approaching waves. (b) Significant wave height field associated with the extratropical transition of Hurricane Danielle (1998). The positions of the buoys are marked. Tight gradients at the leading edge of the maximum significant wave heights are typical for storms undergoing extratropical transition.

phoon Bart from NOGAPS (Fig. 9b) that were initialized during the final tropical cyclone stage (0000 UTC 23 September–0000 UTC 24 September) significantly overdeveloped the extratropical cyclone and forecast the minimum central pressure to occur 12 h too early. Forecasts initiated at 1200 UTC 24 September and beyond were consistently deeper than the analysis during the intensification and weakening stages of the ET.

A second example of the forecast variability associated with ET can be seen in the NOGAPS forecasts of the ET of Typhoon David in 1997 (Fig. 10). During the first 72 h of the forecast sequence initialized at 0000 UTC 16 September (Fig. 10, middle row), David remained a tropical cyclone and both the position and intensity forecasts are quite accurate. Beyond 72 h, when David moved rapidly out over the North Pacific Ocean and underwent ET (Klein et al. 2000), the forecast from 0000 UTC 16 September maintained David as a tropical cyclone that stalled over Japan so that position errors became very large. Although intensity errors appeared to remain small, these forecast intensities were with respect to the tropical cyclone aspect of David and not for the extratropical cyclone that developed downstream. The forecast sequence initialized at 0000 UTC 17 September 1997 (Fig. 10, bottom row) was again quite accurate through 48 h, which was the final tropical stage of David. In this forecast David underwent ET. However, the resulting extratropical system was not forecast to deepen rapidly enough.

A significant contributor to numerical forecast errors during ET is the uncertainty in the initial conditions. Although midocean regions may contain copious amounts of single-level data such as satellite winds (e.g., Velden et al. 1997), there is a lack of conventional multilevel data. Thus a major challenge for numerical forecasts of ET is the optimal use of the available observations. A number of studies have demonstrated the sensitivity of numerical forecasts of ET to the specification of the initial conditions. Evans et al. (2000) obtained an improved forecast of the ET of Hurricane Floyd (1999) by assimilating satellite winds into the initial conditions. Rabier et al. (1996) used an adjoint method based on the 48-h model forecast error to show that changes in the initial conditions could lead to an improved forecast of the explosive extratropical cyclogenesis during the ET of Hurricane Floyd (1993). Hello et al. (2000) demonstrated how variants in the use of the same set of observations could dramatically change the forecast of the ET of Hurricane Iris (1995) over the North Atlantic. Their study suggests that variational data assimilation techniques are only optimal relative to a number of constraints, among others, the specification of background errors. Further refinements will be needed to optimize the handling of observations in unstable environments in which fast error growth can occur (Jarvinen et al. 1999).

Among the most critical assumptions made in data assimilation are the prescriptions of background errors that can be calibrated only by using crude regularity assumptions and a large sample of weather situations (Lorenc 1986; Rabier et al. 1998). Although four-dimensional variational techniques alleviate these constraints by developing flow-dependent, implicit background errors throughout the data assimilation time window, these techniques still have their weaknesses. The assumption that the assimilating model does not have systematic errors means that the impact of observations might be underestimated when they are most needed to correct quickly developing errors. During ET, the tropical characteristics such as warm and humid low-level air masses that exist over a long period of time are a challenge for the convective and boundary layer parameterizations, since inaccuracies in the physical tendencies lead to the low-level humidity being unrealistically

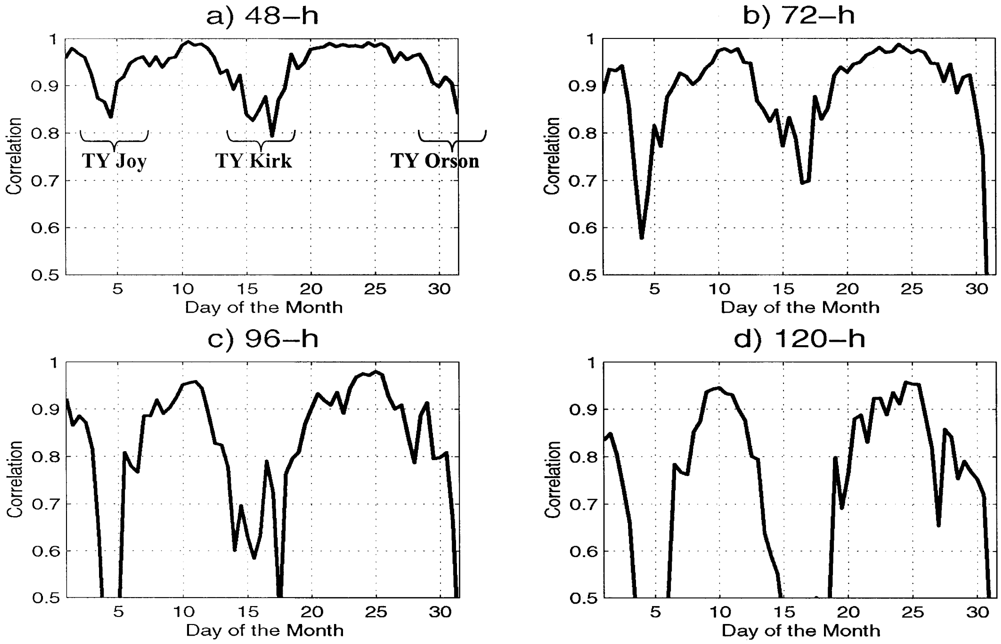

FIG. 8. Anomaly correlations for NOGAPS forecasts of 500-hPa heights over the North Pacific (20°–70°N, 120°E–120°W) during Aug 1996. Each panel represents a specific forecast interval, as labeled. The extratropical transition events that occurred during the month are marked in (a).

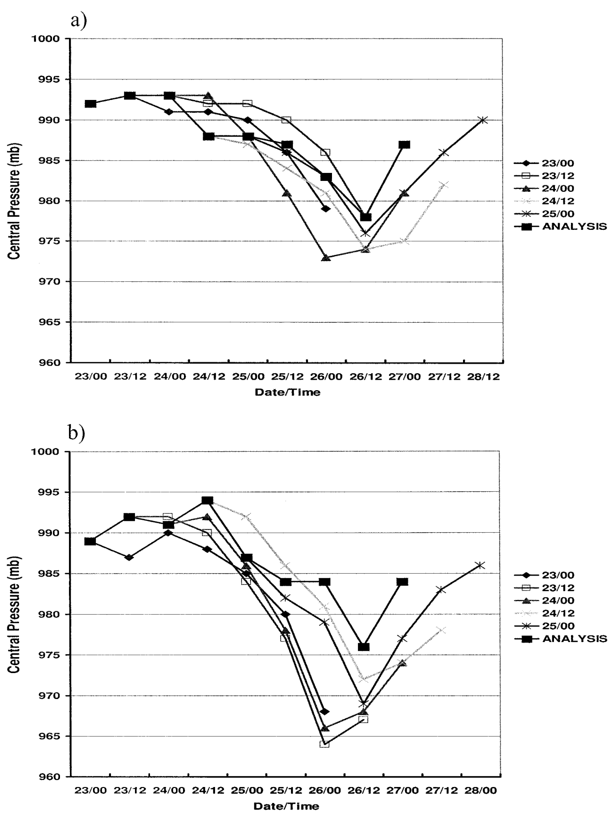

FIG. 9. Analyzed and forecast central sea level pressure for the extratropical transition of Typhoon Bart (1999) from the global model of (a) JMA and (b) NOGAPS. The initialization time of each forecast is shown on the right.

low. The assimilation of humidity data from satellite observations in cloudy environments is one of the most difficult challenges. Indeed, most global numerical weather prediction centers still assimilate clear-air radiances only (McNally and Vesperini 1996). Even when dealing with microwave [Special Sensor Microwave Imager (SSM/I)] data, the spreading of information in the vertical relies almost entirely on the model's representation of the environment (Gérard and Saunders 1999). Finally, there are large uncertainties associated with the coupling between the ocean waves and the atmospheric boundary layer (Eymard et al. 1999). Small changes in the transfer formulations may have a large effect on the heat and momentum fluxes and on the storm development. These challenges are currently being investigated by the different groups involved in numerical weather prediction. New satellite datasets including the Tropical Precipitation Measuring Mission (TRMM) precipitation radar and improved coverage from scatterometers [e.g., the National Aeronautics and Space Administration's Quick Scatterometer (QuikSCAT)] should help locate and diagnose ET in real time.

A further contributor to the uncertainties in the initial conditions is the impact of bogusing the tropical storm vortex. The vortex bogus is implemented in order to improve tropical cyclone track forecasts (DeMaria et al. 1992; Kurihara et al. 1993; Heming and Radford 1998; Davidson and Weber 2000; Pu and Braun 2001). However, the bogus structure is representative of a tropical cyclone and not of an ET event, so that continued implementation of a bogus may lead to the onset of ET being delayed in the numerical forecast.

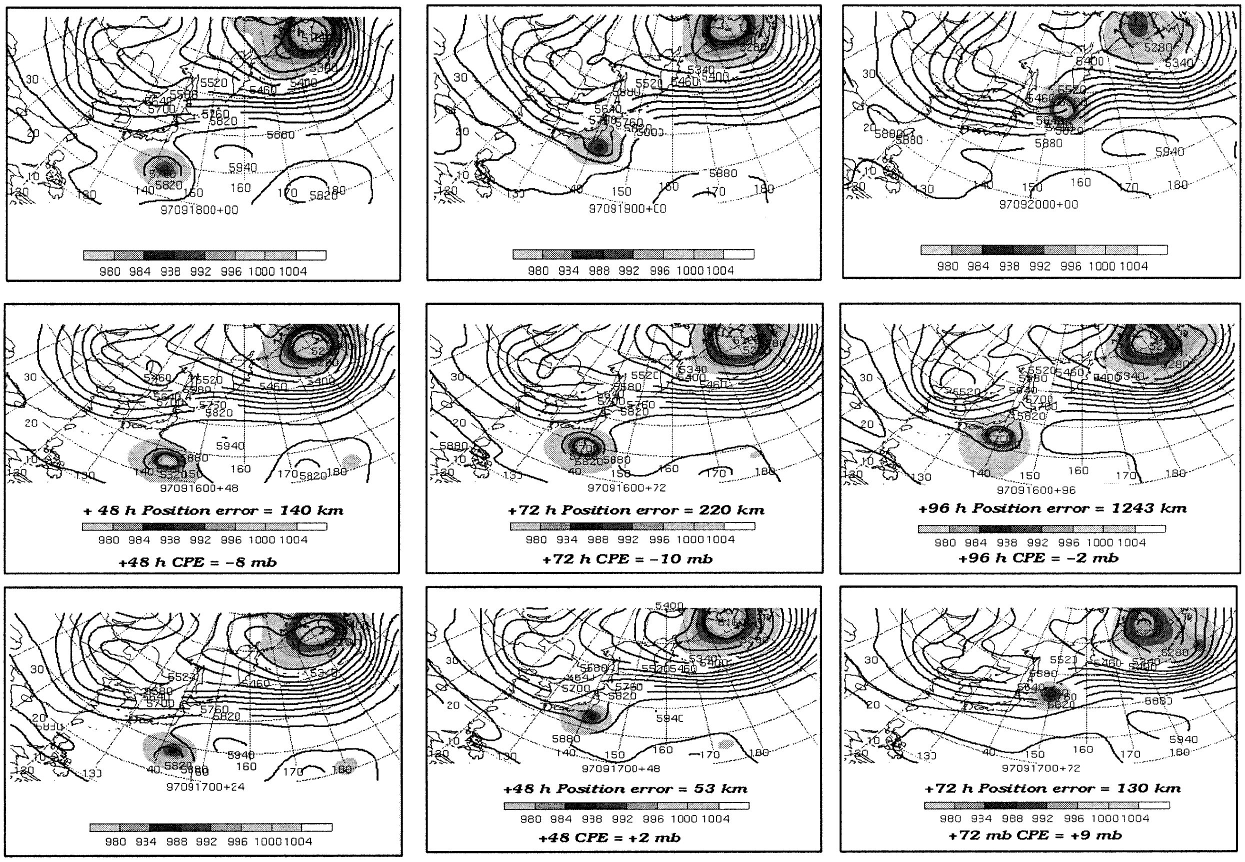

FIG. 10. The 500-hPa height (contours) and mean sea level pressure (shaded) from the NOGAPS model for Typhoon David (1997). Top row: analyses at 0000 UTC 18 Sep (left), 0000 UTC 19 Sep (middle), and 0000 UTC 20 Sep (right). Middle row: corresponding forecasts initialized at 0000 UTC 16 Sep. Bottom row: corresponding forecasts initialized at 0000 UTC 17 Sep.

In order to provide the necessary warning time for a tropical cyclone that is about to undergo ET and move rapidly into the midlatitudes, it is essential to achieve improved accuracy of numerical forecasts in the medium range. Thus, it is necessary to fully diagnose model tendencies and potential sources of model error during the ET of a tropical cyclone. Such a diagnosis requires both increased understanding of the physical aspects of ET and identification of sensitivities to the initial conditions and physical parameterizations of the forecast model. Techniques such as the systematic approach developed for tropical cyclone forecasting (Carr and Elsberry 1994) might provide improved guidance to the forecaster during an ET event. The application of ensemble techniques to ET could improve the value of the ET forecasts. The ensembles currently run in a number of numerical weather prediction centers usually do not explicitly address the uncertainties in a tropical environment. Some extensions of the singular vector technique (Barkmeijer et al. 2001; Puri et al. 2001) that are currently being investigated might significantly improve the handling of ET. Browning et al. (2000) applied singular vector techniques to diagnose the forecast error during the ET of Hurricane Lili (1996) and showed that a comparison of water vapor imagery with model-generated imagery could be used to diagnose the position error in the numerical model of a mesoscale tropopause depression that played an important role in the reintensification. In addition, multimodel ensembles, such as have been applied to tropical cyclone forecasting (Krishnamurti et al. 2000; Weber 2003), may prove useful in the forecast of ET.

3. Current physical understanding of extratropical transition

a. Classifications of extratropical transition

A number of different studies have attempted to classify the evolution of an ET event. Early case studies of ET in the northwest Pacific used surface analyses to classify ET as *complex* when the tropical cyclone interacted with a surface baroclinic zone and *compound* when it interacted with a surface low pressure system (Sekioka 1956, 1970, 1972a,b; Matano and Sekioka 1971a,b; Mohr 1971; Brand and Guard 1979). A third group in this classification occurs when the tropical cyclone remnants dissipate while moving into the midlatitude environment. This classification is used operationally by the JMA (Kitabatake 2002). Foley and Hanstrum (1994) defined two types of ET over the southeast Indian Ocean, which they called cradled, when environmental easterly flow persisted to the south of the cyclone during ET, or captured, when the cyclone became embedded in westerly flow ahead of a cold front.

In a study of three hybrid cyclones, Beven (1997) suggested a cyclone classification based on the temperature anomaly in the cyclone core and the frontal structure. Beven's suggestion was expanded, refined, and quantified into an objective cyclone phase space by Hart (2003). The cyclone phase space is based on two parameters. One parameter describes the magnitude of asymmetries in the thermal structure, using the 900–600-hPa thickness asymmetry centered on the storm track. The other parameter is the vertical derivative of the maximum geopotential height gradient within 500 km of the cyclone center, that, under the assumption of thermal wind balance, indicates whether a cyclone has a warm core, a cold core, or has no thermal anomaly. This parameter is evaluated in the lower and upper troposphere, giving information on the vertical structure of the cyclone. Thus it is possible to distinguish subtropical storms (with a shallow warm core overlaid by a cold core) from tropical or extratropical systems. Since the parameters can be calculated from the three-dimensional height field, the three-dimensional phase space can be obtained from numerical analyses and forecasts. Evans and Hart (2003) use a threshold value of the thermal asymmetry parameter to define the onset of transition and define completion of transition to occur when the cyclone has a cold core in the lower troposphere.

Klein et al. (2000) examined ET events over the western North Pacific using infrared satellite imagery and found that nearly all cases appeared to transform from a warm-core vortex into a baroclinic, extratropical cyclone in a similar manner. They labeled this stage of ET the *transformation* stage. A second stage, labeled the reintensification stage, was added to define the completion of ET as a mature extratropical cyclone.

Extratropical transition can be assessed explicitly by means of classic synoptic tools such as thermal vorticity, the gradient of thermal vorticity, and the advection of absolute vorticity by the thermal wind, all ideas based on Sutcliffe (1939, 1947) and Sutcliffe and Forsdyke (1950). Darr (2002a–c) quantitatively analyzed ET in the Atlantic basin using this approach. He showed that as tropical cyclones move into the midlatitudes and develop characteristics of extratropical cyclones, a positive–negative couplet of ''vorticity advection by the thermal wind'' becomes apparent. The maximum value of this vorticity advection couplet during cyclone recurvature provides a first-order analysis of whether a tropical cyclone will dissipate without becoming extratropical, will undergo ET without significant reintensification, or will reintensify significantly as an extratropical cyclone. The time period of ET is computed by comparing the values of the three Sutcliffe variables to midlatitude thresholds for each Sutcliffe variable. This ET time generally incorporates the reintensification stage, similar to the definition of ET provided by Klein et al. (2000). The onset of ET compares very well also with the phase space methodology of Hart (2003). Finally, in a qualitative sense, ET can be diagnosed by tracking the position of the thermal vorticity maximum (cold-core center) with respect to the low-level circulation center using a radius–azimuth plot. Sanders (1986a) employed this methodology to diagnose explosive cyclogenesis events in the western North Atlantic Ocean by tracking 500-hPa absolute vorticity centers. The rapid cyclogenesis events compare well with significant ET events, with the 500-hPa absolute vorticity maximum (thermal vorticity maximum) moving southwest of the low-level circulation center indicating a strong baroclinic-type development.

Although several classifications have been proposed for a typical ET event, a generalized two-stage process can be identified (Fig. 11). In the remainder of this section we give an overview of a two-stage classification of ET, a slightly modified version of Klein et al. (2000), and introduce some of the processes that play a role in ET.

During the transformation stage, the tropical cyclone begins to respond to the changes in its environment summarized in Fig. 11. This response frequently includes a significant increase in its translation speed (Fig. 3). The structure of the tropical cyclone changes from a nearly symmetric distribution of clouds (Fig. 12a) and circular enclosed eye (Fig. 12b) to contain asymmetries in the wind and thermal structures and in the cloud and precipitation fields (Figs. 12c,d). The convection may become asymmetric in response to vertical shear, moisture variations, SST gradients, or orography. Asymmetries in the surface fluxes may occur as a response to asymmetries in the wind and moisture fields, and to SST gradients. An upper-level jet streak may be seen poleward and to the east of the storm, with an area of warm frontogenesis (Fig. 12e) in the troposphere (Klein et al. 2000; Sinclair 1993b, 2002). Based on satellite imagery, Klein et al. (2000) observed that the combination of the above factors is manifested in a dry slot (Fig. 12c) that modified the symmetric appearance of the tropical cyclone and eroded a portion of the eyewall. Similar to the study of Shimazu (1998), Klein et al. (2000) identified a ''delta rain'' region (Figs. 12d,f) that results from the interaction between the asymmetric decaying tropical cyclone and a midlatitude baroclinic zone. They used the appearance of the dry slot in satellite imagery to define the start of the transformation stage. Evans and Hart (2003) use the terminology ''onset of transition'' rather than ''start of transformation.'' However, the definitions of both Klein et al. (2000) and Evans and Hart (2003) are based on the increasing asymmetry of the tropical cyclone so that there does not appear to be any conflict between these two classifi-

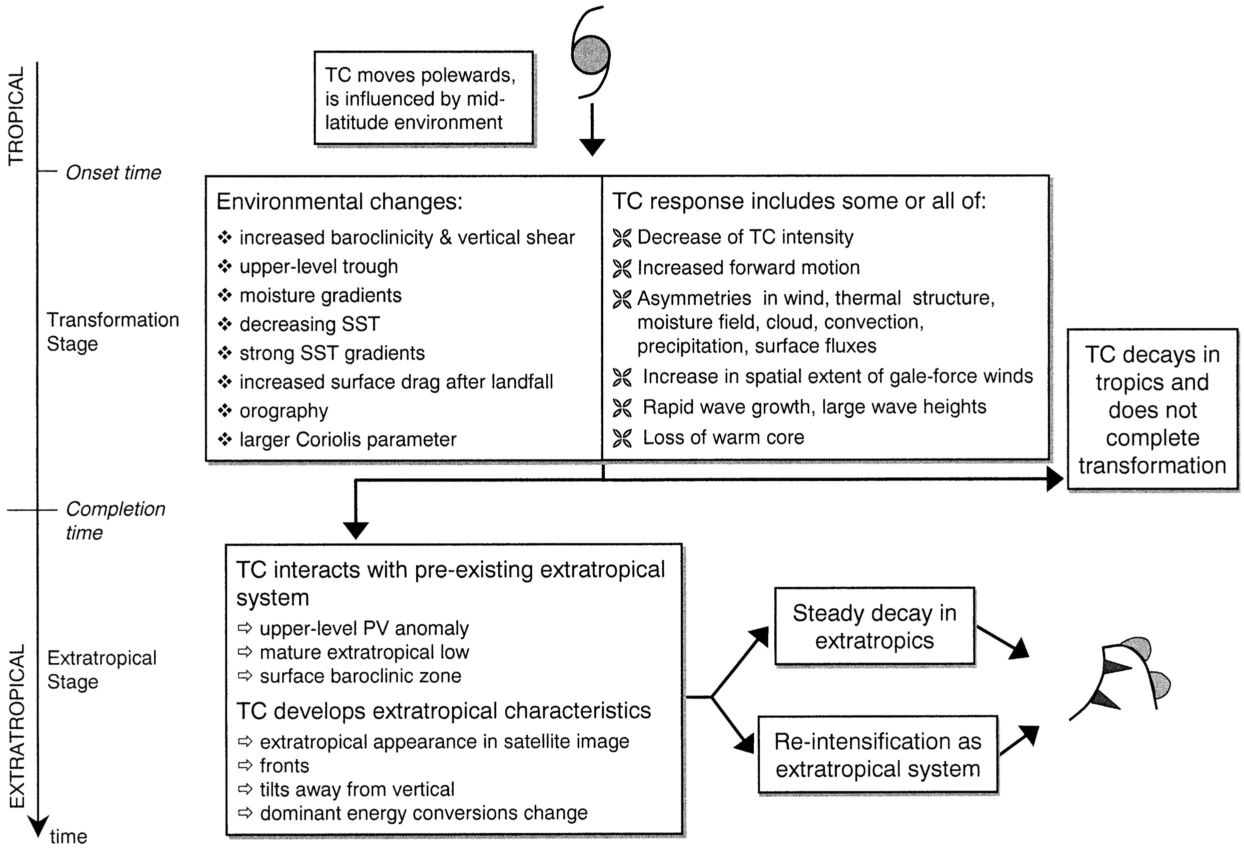

FIG. 11. A two-stage classification of extratropical transition based on the classification of Klein et al. (2000). The onset and completion times correspond to the definitions of Evans and Hart (2003). The "tropical" and "extratropical" labels indicate approximately how the system would be regarded by an operational forecast center.

cations. Rather, two different methods are available to the forecaster to assess the start of the extratropical transition process, one based on satellite imagery and the other based on model analyses. Not all tropical cyclones enter the transformation stage, since extreme environmental factors in the Tropics (vertical wind shear, landfall) may result in the dissipation of a tropical cyclone without the circulation entering the transformation stage of ET. Neither do all tropical cyclones that enter transformation also complete transformation, since factors such as strong vertical shear (section 3c) or reduced SST (section 3f) may lead to the decay of a tropical cyclone during transformation. The completion of the transformation stage should correspond roughly to the time at which operational forecast centers consider the system to be extratropical. Evans and Hart (2003) showed that in the majority of North Atlantic ET cases there is reasonable correspondence between the National Hurricane Center definition of the time of ET and the time at which the system develops a cold-core structure in the lower troposphere.

During the extratropical stage (Fig. 11) the system takes on a more extratropical appearance in satellite imagery, typified by the change from a symmetric region of convection around a cloud-free eye (Fig. 12a) to an asymmetric cloud distribution (Fig. 12g). It may develop extratropical characteristics such as increasing frontogenesis (section 3g), a cold core (Hart 2003; Evans and Hart 2003), and increased conversion of available potential energy to eddy kinetic energy (section 3i). Latent heat release may remain an important source of energy (section 3e) and energy fluxes from the sea surface may continue to play an important role (section 3f). In the extratropical stage the system may continue to decay while developing extratropical characteristics, or may reintensify, in which case Klein et al. (2000) label this as a reintensification stage. However, even if the extratropical system continues to decay, it may still constitute a considerable hazard. Thus we prefer to use the term extratropical stage. During this stage, the extratropical cyclone develops an extensive warm frontal region while development of a well-defined cold front is often suppressed (Fig. 12g) due to a direct thermal circulation that includes the descent of cold air from upstream of the reintensifying cyclone. Either reintensification or decay can occur as the tropical cyclone interacts with a preexisting extratropical system (section 3d). In the absence of a preexisting extratropical system the ex–tropical cyclone would be expected to continue to decay,

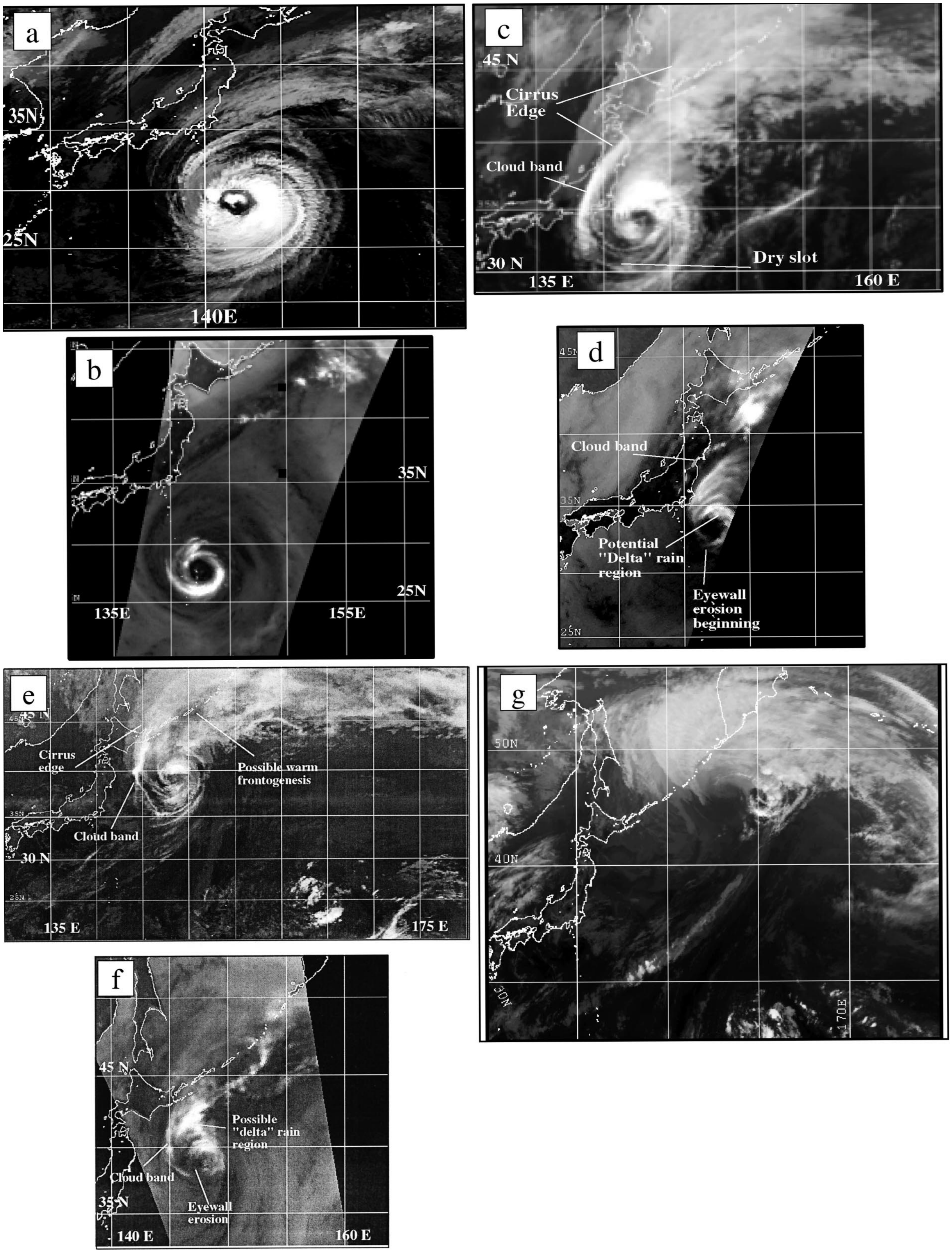

FIG. 12. (a) Geostationery Meteorological Satellite (Japan; GMS) IR image at 2332 UTC 17 Sep 1997 and (b) 85-GHz SSM/I image at 2259 UTC 17 Sep 1997 prior to the start of the extratropical transition of Typhoon David. (c) GMS IR image at 1232 UTC 18 Sep 1997 and (d) 85-GHz SSM/I image at 1120 UTC 18 Sep 1997 during the transformation stage of Typhoon David (1997). Key structural characteristics

although it may still remain a significant low pressure system in the extratropics for some days.

The cyclone phase space of Hart (2003) illustrates some of the changes described above. The path followed through the phase space by a classical ET event is illustrated in Figs. 13a,b for the case of Hurricane Floyd (1999). In the lower troposphere (Fig. 13a) the tropical cyclone has a symmetric, warm-core structure and thus is located in the bottom-right quadrant of the diagram. At the onset of ET the cyclone develops thermal asymmetries, but remains warm cored and moves into the top-right quadrant. The development of an extratropical structure is seen when the system develops a cold core in the lower troposphere and moves into the top-left quadrant. The cold-core structure develops first at upper levels (Fig. 13b). As discussed by Evans and Hart (2003), not all ET events follow this typical path, since in some cases a warm core is retained in the lower troposphere throughout ET [e.g., Hurricane Iris in 1995 (Thorncroft and Jones 2000); Hurricane Lili in 1996 (Browning et al. 1998)]. Hurricane Gabrielle (2001) retained its warm core in the lower troposphere for many days after it developed a thermally asymmetric structure (Fig. 13c), although it had a cold-core structure in the upper troposphere (Fig. 13d). In such a case a parameter based on the existence of a lower-tropospheric cold core does not reflect the operational definition of the time of ET (Evans and Hart 2003).

b. PV thinking—An aid to understanding extratropical transition

The potential vorticity (PV) is a variable that incorporates both dynamic and thermodynamic properties of the atmosphere (Rossby 1940; Ertel 1942). Potential vorticity is conserved in the absence of diabatic and frictional processes (Ertel 1942), and, given a balance condition, a reference state, and suitable boundary conditions, the PV distribution can be inverted to obtain the full wind and mass fields (Kleinschmidt 1950a, 1957; Thorpe 1986). The invertibility of PV enables us to build up a conceptual image of the flow and temperature structure associated with a given PV anomaly. The conservation of PV means that given the PV and wind fields on an isentropic surface we can deduce how the PV anomalies might evolve with time. This so-called PV thinking (Hoskins et al. 1985) can give us a better understanding of the evolution of both tropical and extratropical cyclones. Even in the presence of diabatic or frictional processes it is possible to deduce their contributions to the evolution of the PV field (Ertel 1942; Kleinschmidt 1950b; Cooper et al. 1992).

The application of PV thinking to ET requires a knowledge of the nature of the PV anomalies involved. There are not many detailed observations of the PV structure of a tropical cyclone. One exception is the study of Hurricane Gloria (1985) by Shapiro and Franklin (1995). From this study, and from modeling studies and theoretical considerations, we can gain a broad idea of the PV structure of a tropical cyclone. A tropical cyclone is characterized by strong cyclonic flow throughout the troposphere, with weak anticyclonic flow at larger radii just below the tropopause. The PV structure of the cyclonic part of the tropical cyclone vortex consists of a strong deep positive PV anomaly (relative to the environmental PV) at small radii. The positive PV anomaly is typically surrounded by a much weaker or negative PV anomaly in the lower and middle troposphere, which extends out to large radii. The upper-level anticyclone is characterized by a broad but shallow negative PV anomaly at larger radii. Observations of tropical cyclones show a broad range of structures and intensities suggesting that the amplitude and length scales of the PV anomalies will vary considerably from storm to storm. The PV structure of extratropical weather systems is much better documented. Potential vorticity anomalies are associated with undulations of the tropopause or perturbations on a surface baroclinic zone (e.g., Hoskins et al. 1985) and can arise also through diabatic or frictional processes (e.g., Thorpe and Clough 1991; Davis and Emanuel 1991; Hoskins et al. 1985).

The technique of piecewise PV inversion (Robinson 1988; Davis 1992), whereby the flow and mass fields associated with a particular portion of the PV field are calculated, can be used to deduce which PV features play an important role in cyclogenesis. The use of piecewise PV inversion has provided significant insights into the dynamics of extratropical cyclones (e.g., Thorpe 1987; Davis and Emanuel 1991; Davis 1992; Davis et al. 1993, 1996; Gyakum et al. 1995; Stoelinga 1996; Huo et al. 1998; Morgan 1999) and tropical cyclones (e.g., Wu and Emanuel 1995a,b; Shapiro 1996; Wu and Kurihara 1996; Möller and Jones 1998; Henderson et al. 1999; Shapiro and Franklin 1999; Möller and Shapiro 2002; Shapiro and Möller 2003). More recently piecewise PV inversion has been used to assess the importance of particular features, such as an upper-level trough or the tropical cyclone, during ET (Browning et al. 2000; McTaggart-Cowan et al. 2001, 2003b; APQJR) and to assess the relative contributions of dynamics and thermodynamics to cyclogenesis (McTaggart-Cowan et al. 2003a).

In order to perform piecewise PV inversion for a

←

are labeled. (e) GMS IR image at 1232 UTC 19 Sep 1997 and (f) 85-GHz SSM/I image at 1120 UTC 19 Sep 1997 during the start of the extratropical stage of Typhoon David (1997) (g) Infrared GMS imagery of the extratropical stage of Typhoon David at 0032 UTC 20 Sep 1997. [Adapted from Klein et al. (2000).]

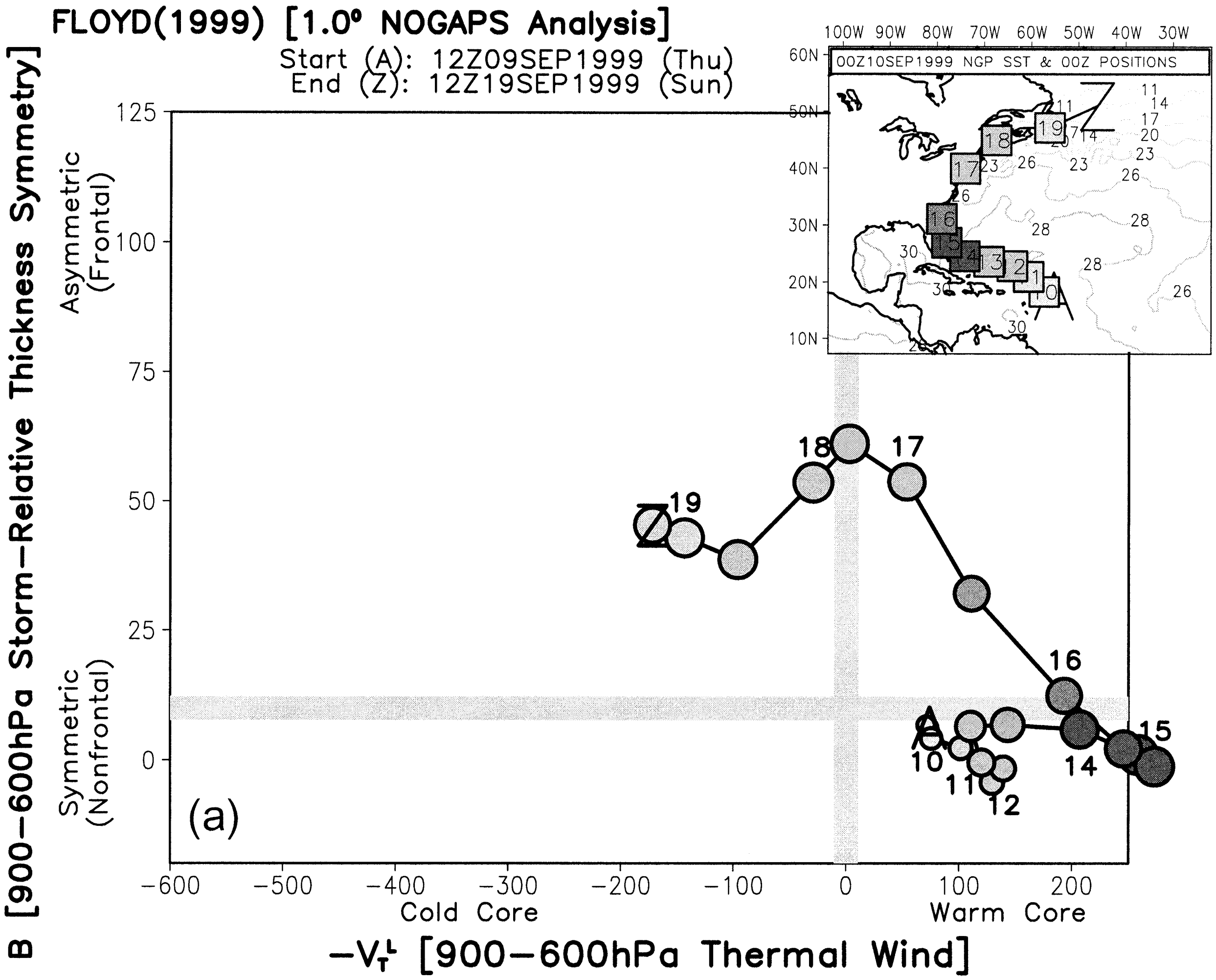

FIG. 13. Cyclone phase space diagrams for (a), (b) the extratropical transition of Hurricane Floyd in 1999 and (c), (d) the extratropical transition of Hurricane Gabrielle in 2001. In all diagrams the abscissa indicates whether the cyclone is warm or cold core in the lower troposphere (strong cold core on left; strong warm core on right). In (a), (c) the ordinate indicates the degree of thermal asymmetry in the lower troposphere (symmetric at bottom; asymmetric at top), In (b), (d) the ordinate displays the cold- /warm-core structure of the upper troposphere (cold core at bottom; warm core at top). Here, A indicates the first time, and Z the final time plotted. Markers are plotted every 12 h with the size of the marker indicating the mean radius of the 925-hPa gale force winds and the shading indicating the minimum mean sea level pressure (>1010 hPa white; <970 hPa black). The inset gives the track of the cyclones superposed on the model analysis SST field (°C). [(a), (b) Taken from Hart (2003), (c) from Evans and Hart (2003), and (d) from http://eyewall.met.psu.edu/cyclonephase.]

nonlinear system such as an ET event, the choice of a background state and a balance condition must be addressed. The Rossby number in an ET event would be expected to be ≫1, so that formally quasigeostrophic theory would not apply. However, it has been shown that quasigeostrophic theory can be used to gain insight into cyclogenesis events, even when it is not formally valid (Hakim et al. 1996). A number of studies have used a climatological background state and an ad hoc linearization of the nonlinear balance equation to study extratropical and tropical cyclones (Davis and Emanuel 1991; Wu and Emanuel 1995a,b; Wu and Kurihara 1996). When piecewise PV inversion is used for a tropical cyclone, it is advantageous to use a symmetric vortex as the basic state (Shapiro 1996; Shapiro and Franklin 1999), since the asymmetries in a tropical cyclone are typically much weaker than the symmetric flow (Shapiro and Montgomery 1993). An alternative balance condition to the nonlinear balance equations is the asymmetric balance theory of Shapiro and Montgomery (1993), which has the advantage of allowing for divergence that is not small. However, although this technique has advantages for the study of tropical cyclones, no investigations have yet considered possible applications to ET when the asymmetric flow is larger in amplitude.

The application of PV thinking to ET presents an observational challenge. In order to compute both dry and moist PV properly it is necessary to map the time-dependent three-dimensional structure of the wind, tem-

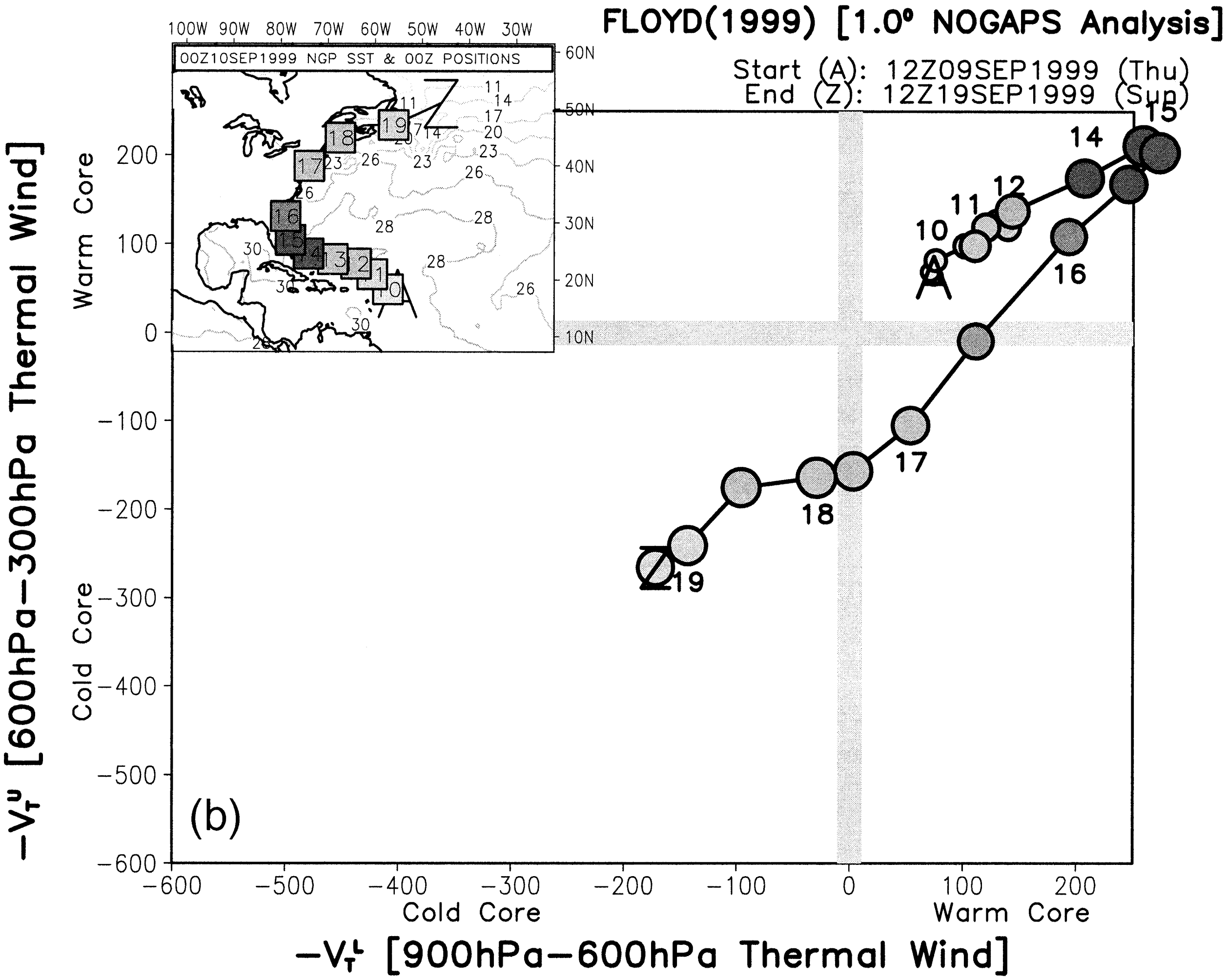

FIG. 13. (*Continued*)

perature, and moisture fields of both the storm and its environment.

c. Interaction with environmental vertical shear during the transformation phase

A better understanding of the extratropical development that occurs after ET might be gained from the initial value problem for baroclinic instability. The idea that the nature of the initial perturbation to a baroclinically unstable state can influence the subsequent growth has been discussed by many authors. For example, Farrell (1985, 1989a) showed that wavelike perturbations with an orientation against the shear lead to significant growth even when the basic state is stable to normal mode growth. Although these studies suggest the importance of the initial conditions for subsequent baroclinic growth, the wavelike perturbations considered have little resemblance to the structure of a tropical cyclone. However, Badger and Hoskins (2001) showed that the magnitude of baroclinic growth in the Eady model (Eady 1949) is also sensitive to the scale, location, and tilt of more localized perturbations. It is clear therefore that changes in the structure of the tropical cyclone as it undergoes transformation may have a strong influence on subsequent developments in a baroclinically unstable flow.

One factor that can lead to substantial modification of the structure and intensity of a tropical cyclone during the transformation stage is environmental vertical shear of the horizontal wind. Idealized modeling studies of the influence of vertical shear on tropical cyclones have shown that the main structural changes are associated with the development of vertical tilt and of asymmetries in the vertical motion and temperature fields (Jones 1995, 2000a,b; DeMaria 1996; Bender 1997; Frank and Ritchie 1999, 2001). Jones (1995, 2000a,b) showed that the vertical tilt that a tropical cyclone develops in a vertically sheared environment may change with time and must not necessarily be a downshear tilt. The magnitude and direction of the tilt depends on the strength of the shear, the strength and size of the tropical cyclone,

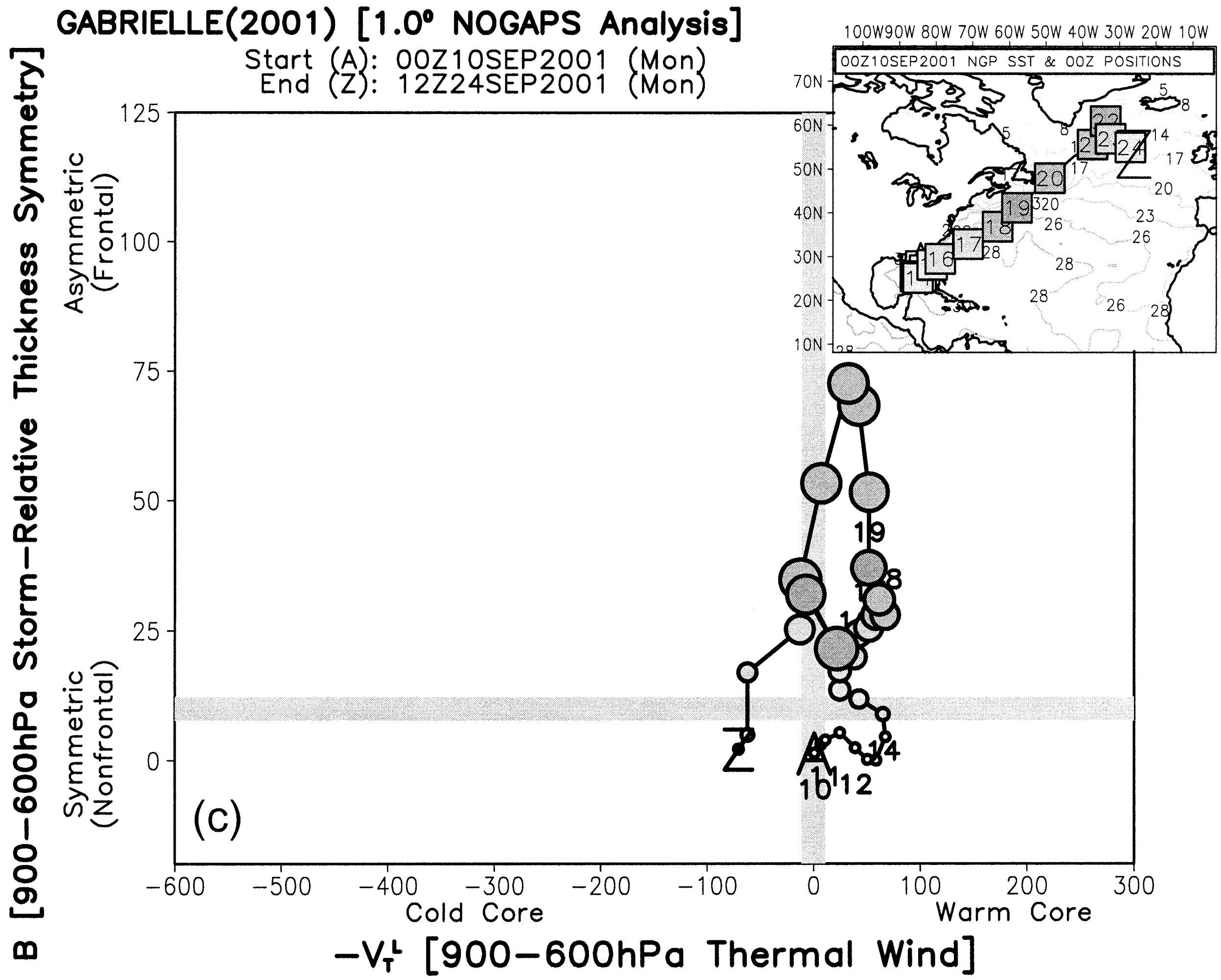

FIG. 13. (*Continued*)

the environmental Brunt–Väisälä frequency, and the Coriolis parameter. In a dry vortex the orientation of the vertical motion and temperature asymmetries are directly linked to the direction of the vertical tilt (Jones 1995, 2000b). Their development can be explained in terms of the maintenance of balance (Raymond 1992; Jones 1995, 2000b). Frank and Ritchie (1999, 2001) showed that in a moist model the pattern of asymmetric vertical motion changes when the inner core of the tropical cyclone becomes saturated. Further work is needed to elucidate the role of these structural changes in ET.

The studies of the interaction of a tropical cyclone with environmental shear used the relatively weak shears typical of the tropical environment [generally less than 2 m s^{-1} (100 hPa)$^{-1}$]. As a tropical cyclone approaches the midlatitude westerlies, the environmental vertical shear will increase. For example, Thorncroft and Jones (2000) calculated a vertical shear of over 5 m s^{-1} (100 hPa)$^{-1}$ during the transformation stage of the ET of Hurricane Iris in 1995. When the exhurricane moved into the region of strong shear, the PV of the hurricane inner core became tilted strongly away from the vertical (Fig. 14). A downshear tilt was also seen in the case studies of Typhoon David (Klein et al. 2000) and Hurricane Irene (APQJR). Ritchie and Elsberry (2001) modeled the interaction of an idealized tropical cyclone with a vertical shear of over 3 m s^{-1} (100 hPa)$^{-1}$ and showed that the vertical extent of the PV anomaly of the tropical cyclone decreased and the lower portion of the PV anomaly became tilted downshear.

The above studies suggest that in the presence of strong vertical shear the upper-level PV is eroded leaving a mid- to lower-tropospheric PV anomaly, which is either vertically stacked or has a downshear tilt. The erosion of the upper-level PV may be due to differential advection by the vertical shear, or due to filaments of PV breaking off from the main PV anomaly at upper levels (Jones 2000b). The westward tilt with height necessary for reintensification as an extratropical cyclone occurs in association with the approach of an upper-level PV anomaly from the west, as seen in a number of case studies (Palmén 1958; Chien and Smith 1977;

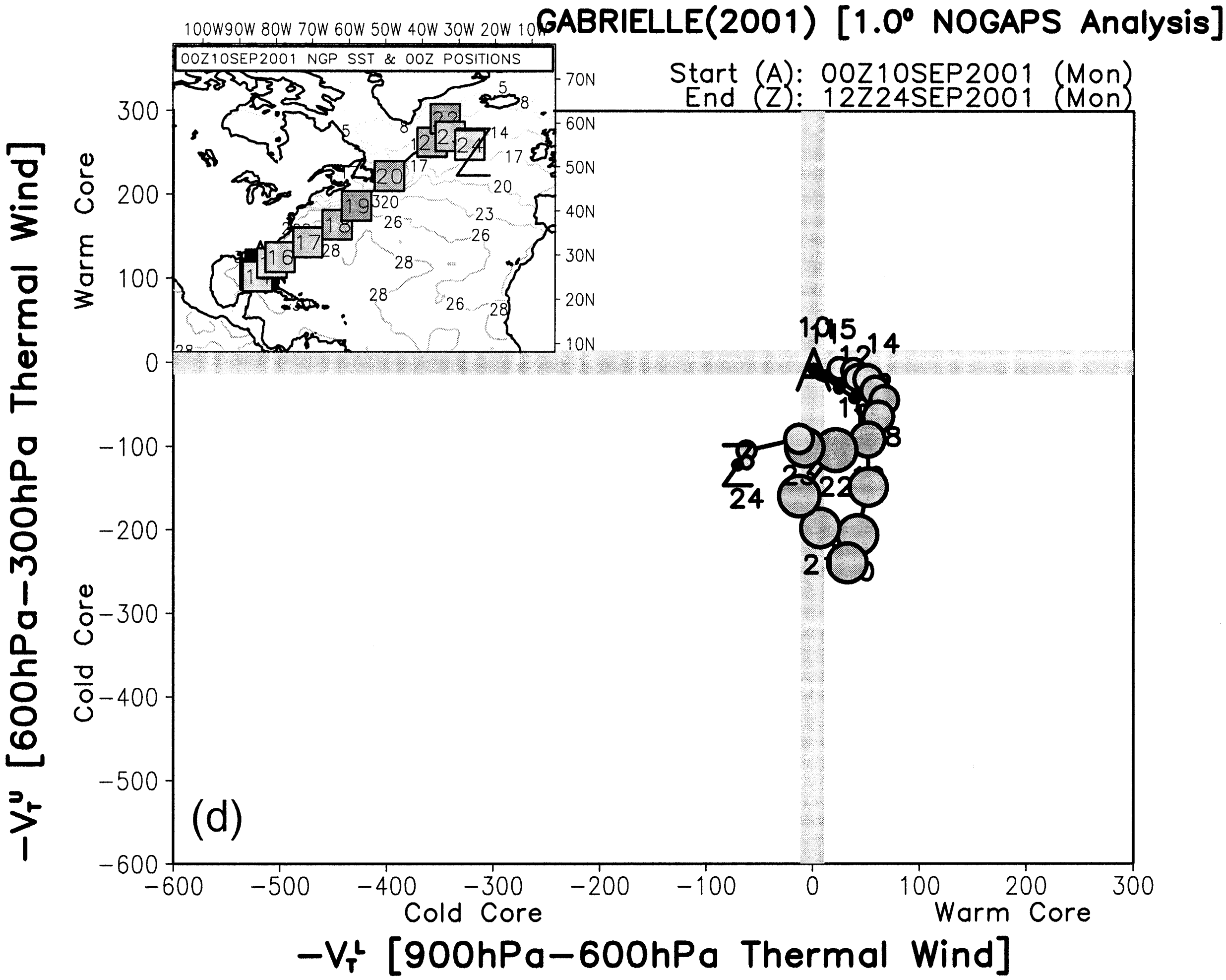

FIG. 13. (*Continued*)

Sinclair 1993b; Bosart and Lackmann 1995; Thorncroft and Jones 2000). This description is consistent with the composite cross section of South Pacific ET in which the relative vorticity is almost vertically stacked in the lower troposphere but there is a strong westward tilt with height in the upper troposphere (Sinclair 2002, Fig. 11). Several days may elapse between the modification of the PV of the tropical cyclone by shear and the interaction with an upper-level PV anomaly [e.g., Patsy in 1986 (Sinclair 1993b)]. More observational and numerical studies are needed to shed light on the precise sequence of events as a tropical cyclone enters the strong vertical shear of the midlatitude westerlies.

d. Midlatitude circulations and trough interaction

Extratropical transition is sensitive to the interaction between the decaying tropical cyclone and the midlatitude circulation into which the tropical cyclone is moving. A recent study based on a large number of cases showed that ET in the western North Pacific is predominantly associated with one of the two synoptic-scale patterns shown in Fig. 15 (Harr and Elsberry 2000; Harr et al. 2000) in which the primary midlatitude circulation is either to the northeast or northwest of the tropical cyclone. The position of the primary midlatitude circulation relative to the poleward-moving tropical cyclone was found to influence frontogenesis and the energy budget during ET (sections 3g,i). A statistically significant relationship between the final central sea level pressure of an ET event and the type of midlatitude circulation pattern into which the tropical cyclone moved was found by Klein et al. (2000). Tropical cyclones that moved into the northwest pattern (Figs. 15a,b) were deeper after 36 h of reintensification than those that moved into the northeast pattern (Figs. 15c,d). Thorncroft et al. (1993) described two paradigms of baroclinic wave life cycle behavior. One of these is dominated by cyclonic Rossby wave breaking and characterized by a cyclonic wrapup of the upper-level PV and a broad and deep surface low; the other is dominated by anticyclonic wave breaking and characterized by the

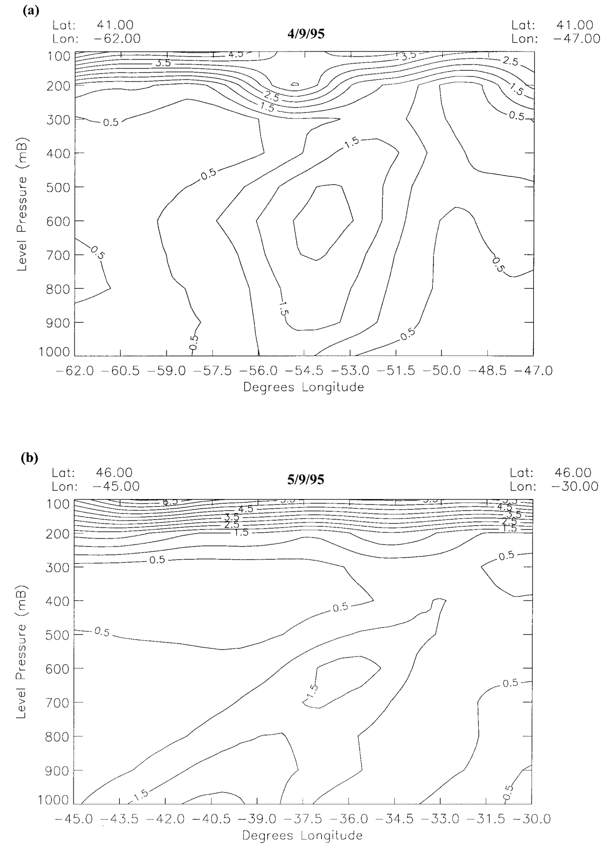

FIG. 14. Vertical sections of Ertel potential vorticity during the extratropical transition of Hurricane Iris in 1995 with a contour interval of 0.5×10^{-6} K m^2 s^{-1} kg^{-1}: (a) 0000 UTC 4 Sep 1995 between 41°N, 62°W and 41°N, 47°W and (b) 0000 UTC 5 Sep 1995 between 46°N, 45°W and 46°N, 30°W. [Taken from Thorncroft and Jones (2000).]

meridional elongation of upper-level PV anomalies and the development of cutoff lows. In the North Atlantic a number of ET events have occurred in association with the cyclonic wave breaking life cycle [Iris in 1995 (Thorncroft and Jones 2000); Lili in 1996 (Browning et al. 1998, 2000); Earl in 1998 (McTaggart-Cowan et al. 2001); Irene in 1999 (APQJR); Gabrielle in 2001 (Evans and Hart 2003)]. In the northwest Pacific ET is more frequently associated with the life cycle in which anticyclonic Rossby wave breaking dominates. An expansion of the study of Klein et al. (2000) should help to determine whether dominant large-scale circulation patterns associated with ET exist in other ocean basins and to identify dissimilarities between ET in different ocean basins.

In the majority of ET events there is an interaction between the decaying tropical cyclone and an upper-level trough. Hanley (1999) constructed composites of Atlantic tropical cyclones that interacted with upper-level troughs and underwent extratropical transition. The ET composite (Fig. 16) contains a set of 14 cases of tropical cyclones that have undergone a trough interaction (as defined by Hanley et al. 2001) and then intensified 10 hPa or more as an extratropical cyclone. In the ET composite, an upper-level PV anomaly approaches the storm center from the northwest (Fig. 16a). In this case, the upper-level PV anomaly is much deeper and wider than in cases when a trough interaction resulted in intensification of the tropical cyclone (Hanley et al. 2001). During the ET process, a region of high PV begins to wrap around the composite center (Figs. 16c,e). Associated with the PV anomaly is very strong vertical shear (up to 24.5 m s^{-1} between 850 and 200 hPa). The shear is observed to weaken in time in the ET composite. The shear in the ET composite is on the order of 2–3 times the shear observed in the case of a trough interaction that results in tropical cyclone intensification. During the early stages of the interaction between the tropical cyclone and the upper-level PV anomaly, the vertical shear will act to decrease the intensity of the tropical cyclone (DeMaria 1996). In the later stages of ET the presence of baroclinity and thus vertical shear is essential for reintensification as an extratropical system. At the start of the period of reintensification as an extratropical cyclone, the composite tropical cyclone is located between two upper-level jets (Fig. 16d). This is a region of enhanced divergence associated with the combination of the upstream jet exit region and the downstream jet entrance region. Uccellini and Kocin (1987) found this to be a favorable region for rapid cyclogenesis in midlatitude systems. The mechanism by which the jet splits between the composite in Fig. 16b and Fig. 16d may be related to the diabatic erosion of the upper-level PV due to strong latent heating in the region downstream of the composite center. A double-jet signature aloft was seen also in the ET of Hurricane Earl in 1998 (McTaggart-Cowan et al. 2001) and in the southwest Pacific composites of Sinclair (2002).

McTaggart-Cowan et al. (2001) used PV inversion techniques to demonstrate that the upper-level trough was crucial for the reintensification of Hurricane Earl in the extratropics, whereas the circulation of the hurricane itself did not play such an important role. In a further study, McTaggart-Cowan et al. (2003b) demonstrated that the ET of Earl was sensitive also to the structure of the downstream flow. They attributed this sensitivity to whether the hurricane remnants were located in the right-entrance or left-exit region of an upper-level jet. The combination of the ageostrophic circulation associated with the jet and the hurricane circulation led to enhanced cold advection to the west of the hurricane in the former case, and enhanced warm advection to the east of the hurricane in the latter case. A further example of the contribution of an upper-level trough to ET is given for Hurricane Floyd (1999) by Atallah and Bosart (2003).

The role played by an upper-level trough has led to some ET events being compared to Petterssen–Smebye type-B cyclogenesis, in which extratropical development occurs when a strong upper-level trough moves

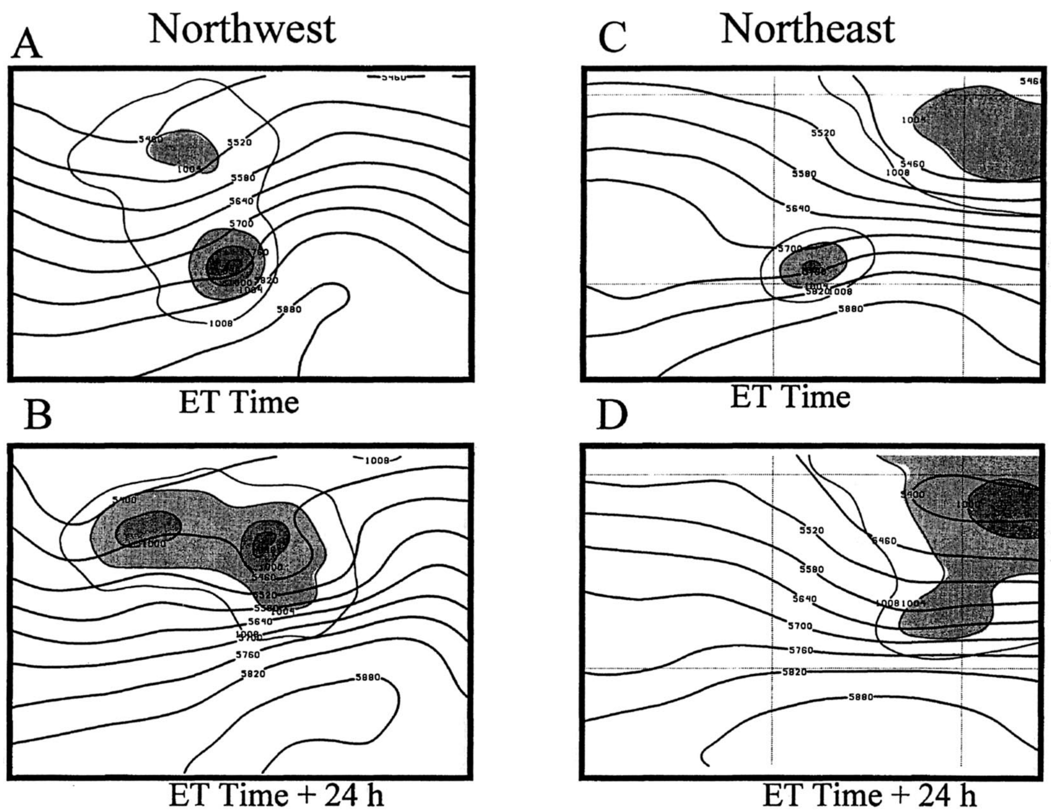

FIG. 15. Composite 500-hPa height (m) and sea level pressure (hPa, only below 1008 hPa with shading in 4-hPa increments starting at 1004 hPa) analyses based on grouping of cases in a northwest pattern at the (a) ET time and (b) ET + 24 h, and in a northeast pattern at the (c) ET time and (d) ET + 24 h. The composite northwest pattern is based on 13 cases and the northeast pattern is based in 17 cases. [Taken from Harr and Elsberry (2000).]

over a region of low-level warm advection (Petterssen and Smebye 1971). During ET, the type-B cyclogenesis is modified by the lower-level remnants of a tropical cyclone and may be enhanced due to the reduced stability associated with the presence of moist tropical air (DiMego and Bosart 1982a,b; Bosart and Dean 1991; Sinclair 1993b). In some cases of ET an upper-level trough may be involved both in the initial development of the tropical cyclone and in the ET (e.g., Hurricane Diana in 1984, Hurricane Michael in 2000). In such cases the tropical cyclone develops out of a baroclinic cyclone that was initiated by an upstream upper-level trough (Bosart and Bartlo 1991; Davis and Bosart 2001, 2002).

Klein et al. (2002) examined the sensitivity of the extratropical stage of ET to the relative interaction between mid- and upper-level midlatitude dynamical processes and low-level thermal processes associated with the decaying tropical cyclone. Reintensification as an extratropical cyclone was found to occur when the combination of the dynamic and thermodynamic processes acted to create a region that was favorable for extratropical cyclone development. The favorable region often develops when the upper-level tropical cyclone outflow enhances the equatorward entrance region of a downstream jet streak and the remnant tropical cyclone circulation interacts with the lower-tropospheric baroclinic zone. Therefore, both tropical cyclone and midlatitude components make significant contributions to reintensification during ET.

It is not always recognized that the interaction between a tropical cyclone and an extratropical system can lead to the tropical cyclone modifying the extratropical circulation. Two mechanisms by which such a modification can occur have been discussed in the literature: the adiabatic interaction between the PV of a tropical cyclone and that of a midlatitude jet, and the impact of diabatically modified PV on the midlatitude circulation.

One simple conceptual model for the first mechanism considers the tropical cyclone interacting with the PV structure of a two-dimensional midlatitude westerly jet. Through action at a distance (Bishop and Thorpe 1994), we expect the circulation associated with the tropical cyclone to excite Rossby waves on the upper-level PV gradients associated with the jet before the PV anomalies of the tropical cyclone and jet themselves meet. The Rossby waves will disperse on the PV gradient in a similar manner to that discussed by Simmons and Hoskins (1979). In their idealized experiments signifi-

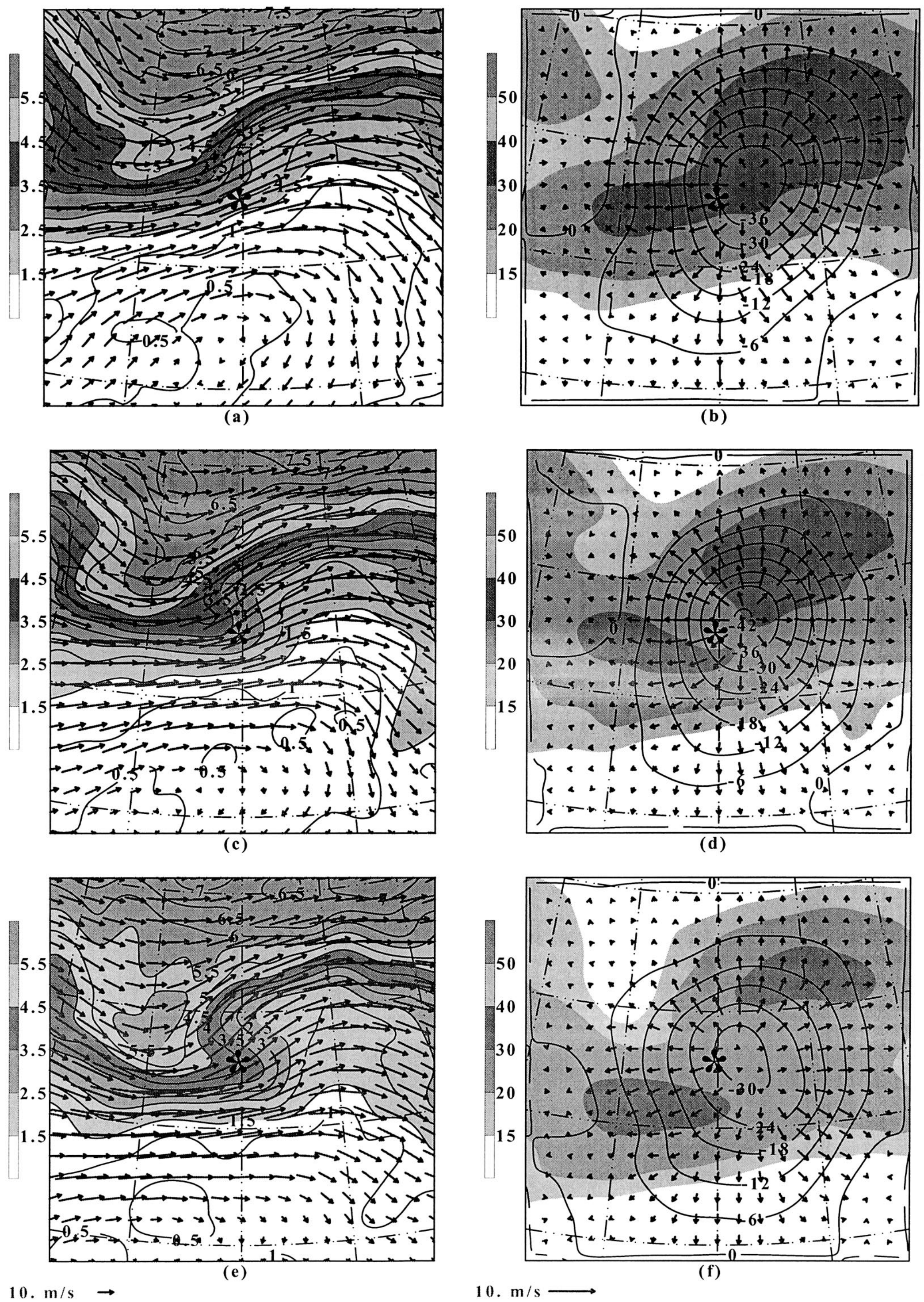

FIG. 16. Horizontal plot on the 200-hPa surface for the ET composite containing cases with a total decrease in central pressure of more than 10 hPa. (a), (c), (e) Vectors of the total wind [reference arrow indicated at the bottom left of (e)], and Ertel potential vorticity (increment of 0.5×10^{-6} K m^2 s^{-1} kg^{-1}; values greater than 1.5×10^{-6} K m^2 s^{-1} kg^{-1} shaded as indicated) at times $t_0 - 12$ h, t_0, and $t_0 + 12$ h, respectively. (b), (d), (f) Total wind speed (values greater than 15 m s^{-1} shaded as indicated), velocity potential (solid lines, contour interval 6×10^5 m^2 s^{-1}), and divergent wind [reference arrow indicated at the bottom left of (f)] at times $t_0 - 12$ h, t_0, and $t_0 + 12$ h, respectively. Here, t_0 is the 12-hourly observation time immediately before the central sea level pressure begins to fall. Asterisk denotes the location of the composite tropical cyclone center, and the increment in latitude and longitude is 10°. [Taken from Hanley (1999).]

cant upper-level developments occur downstream of an initial isolated disturbance illustrating the importance of downstream development for synoptic forecasting. Thus, even if the PV associated with the tropical cyclone does not merge with the midlatitude jet, we expect to see midlatitude developments downstream of the tropical cyclone. Ferreira and Schubert (1999) discussed downstream development in the idealized framework of a shallow-water model. They showed that, through ''action-at-a-distance,'' a cyclonic PV anomaly remote from a midlatitude jet could indeed generate downstream trough development. In addition to downstream development at upper levels, smaller-scale developments can take place upstream at lower levels in association with disturbances growing on the low-level temperature gradient (Simmons and Hoskins 1979; Thorncroft and Hoskins 1990; Wernli et al. 1999). The fact that the circulation of the tropical cyclone has maximum amplitude at low levels suggests that upstream development may be important in some cases.

The second mechanism concerns the modification of the upper-level structure in the extratropics by low PV originating in the outflow of a tropical cyclone. In regions of convection and associated latent heat release, such as in the inner core of a tropical cyclone, the PV is no longer conserved (e.g., Ertel 1942; Hoskins et al. 1985; Raymond 1992). The diabatic modification of PV in regions of latent heat release leads to enhanced PV in the lower and midtroposphere (below the level of maximum heating) and reduced PV in the upper troposphere (Kleinschmidt 1950b; see also Thorpe 1993; Wernli and Davies 1997). A number of recent studies have described how the diabatic reduction of PV in the upper troposphere can result in enhanced ridging downstream of the system in which the latent heat release occurs (Davis et al. 1993; Stoelinga 1996; Dickinson et al. 1997; Bosart 1999, Fig. 9; Henderson et al. 1999; Massacand et al. 2001). This feature is often underestimated in NWP models (Henderson et al. 1999; Atallah and Bosart 2003, Fig. 15). The enhanced ridging may be viewed also as a steepening of the tropopause. Thorncroft and Jones (2000) suggested that the development of the baroclinic life cycle dominated by anticyclonic wave breaking (Davies et al. 1991; Thorncroft et al. 1993) might be promoted downstream of the tropical cyclone by the low-PV tropical cyclone outflow. In the case of Hurricane Floyd (1999) the diabatic outflow of the tropical cyclone assisted in the formation of a tropopause fold (Atallah and Bosart 2003).

The presence of low-PV air in the upper troposphere has been shown to have a significant impact on extratropical cyclogenesis (Hoskins and Berrisford 1988; Pomroy and Thorpe 2000). The low-PV outflow certainly had a strong impact on the ET of Hurricane David in 1979 (Bosart and Lackmann 1995) as there was no significant upper-level PV anomaly apparent until the outflow from the hurricane modified the tropopause structure to the west of the tropical cyclone. The diabatically induced tropopause anomaly subsequently interacted with David resulting in significant development as an extratropical system. A further impact on ET might occur if the flow associated with the low-PV outflow contributes to the steering flow of the tropical cyclone (Henderson et al. 1999).

If the interactions between a tropical cyclone and the midlatitude PV gradients at upper levels are misrepresented in a forecast model, there is likely to be an impact on developments downstream and thus a reduction in skill of medium-range forecasts. Two examples of this are the October storm of 1987 (Hoskins and Berrisford 1988) and the reduced skill of the European Center for Medium-Range Weather Forecasts (ECMWF) forecast over North America in May 2000. Errors in the latter forecast were traced back to Tropical Storm Cimaron in the northwest Pacific. In addition, the changes in the midlatitude circulation pattern due to the presence of the tropical cyclone may affect the track of a tropical cyclone, as shown by Carr and Elsberry (2000), and thus feedback onto the evolution of the ET event.

e. Latent heat release and convection

The release of latent heat both in dynamically forced saturated ascent and in deep convection can be expected to play an important role in ET, both during the transformation and the extratropical stages. During the transformation stage the decrease in intensity of the tropical cyclone will depend on how the inner-core convection is modified by the environmental changes (e.g., decreased SST, increased vertical shear, landfall) experienced by the tropical cyclone. In turn, this modification will affect the structure of the subsequent extratropical system and thus the behavior of this system in the extratropics. In the extratropical stage moist processes may be crucial to the reintensification as an extratropical system.

The importance of latent heat release during ET has been demonstrated in several studies. Edmon and Vincent (1976) compared model simulations of the extratropical transition of Tropical Storm Candy (1976) to an extratropical system over the continental United States using two cumulus parameterization schemes, but found neither scheme to be particularly good in capturing the observed rainfall distribution produced by the storm. However, their study did emphasize the problem of using a parameterization scheme over land in the middle latitudes that was developed for the Tropics. Anthes (1990) compared simulations of the metamorphosis of Hurricane Hazel (1954) with and without latent heat release. The calculations pointed to the great importance of latent heat release in the transition process, but the mechanisms involved were not discussed. Prater and Evans (2002) showed that changing the cumulus parameterization scheme in a numerical simulation of the ET of Hurricane Irene (1999) led to reintensification occurring after ET with the Kain–Fritsch convection scheme (Kain and Fritsch 1990), but not with the Betts–Miller scheme (Betts and Miller 1986). The two simulations began to differ in the tropical cyclone phase, when the maximum of convective heating in the tropical cyclone occurred at different heights for the two schemes, leading to a deeper PV anomaly in one of the cases and thus a different steering flow. As a result of the different steering flow, the tropical cyclone in the Kain–Fritsch case interacted strongly with an upper-level trough leading to reintensification after ET, whereas that in the Betts–Miller case decayed in the central Atlantic.

In terms of PV thinking the impact of latent heat release on a cyclone depends on the gradient of diabatic heating along the absolute vorticity vector (Hoskins et al. 1985; Raymond 1992). In a tropical cyclone the vertical component of absolute vorticity is likely to be dominant in the lower troposphere, so that the vertical gradient of heating plays an important role there, whereas in an extratropical cyclone the absolute vorticity vector will be more strongly tilted away from the vertical. Further characteristics of convection and diabatic heating in a tropical and an extratropical cyclone are discussed below (see section 3d for a discussion of the impact of the diabatically modified upper-level negative PV anomaly).

A useful way of thinking about the effects of convective processes in a tropical cyclone was developed in a series of papers by Emanuel (1991 and references therein, 1995b). Supported by observational studies of convecting maritime atmospheres (e.g., Betts 1982; Xu and Emanuel 1989), deep convection is assumed to bring the free troposphere to a state that is closely approximated locally by a saturation moist adiabat with equivalent potential temperature equal to the equivalent potential temperature (θ_e) at the top of the subcloud layer. The implication is that the virtual temperature, and hence the horizontal density gradient, at any pressure level in the troposphere are determined by the distribution of θ_e in the subcloud layer. In essence, convection is regarded as a type of mixing process that brings the atmosphere to a thermodynamic state that is closely related to local conditions in the subcloud layer (see Emanuel et al. 1994). While the accuracy of this picture is still a matter of controversy (see Stevens et al. 1997; Emanuel et al. 1997), the ideas are useful for qualitative understanding. In fact, as a tropical cyclone matures, the eyewall cloud appears to take on more of the character of slantwise moist ascent (Ooyama 1982; Willoughby 1995 and references therein; Zhu et al. 2001), rather than buoyant convection in the usual sense. The slantwise moist ascent is part of a secondary (radius–height) circulation associated with the unbalanced part of the radial density gradient and with frictional convergence in the boundary layer (Smith 2000).

Using Emanuel's arguments, changes in the radial distribution of boundary layer θ_e in a tropical cyclone lead to changes in the radial density gradient throughout the troposphere. These changes must be accompanied by a change in the secondary circulation to maintain hydrostatic and gradient wind balance, leading to a change in intensity of the tropical cyclone. Thus the problem of understanding the role of moist processes in a tropical cyclone is reduced to understanding and quantifying the processes that determine the distribution of boundary layer θ_e. There are three processes involved in determining the radial gradient of boundary layer θ_e: the surface flux of moist entropy, which is an increasing function of both surface wind speed and falling pressure; the downward flux of air with low entropy through the top of the subcloud layer associated with clear-air subsidence and precipitation-cooled downdrafts; and the horizontal convergence of entropy (Raymond 1995, 1997; Emanuel 1995b).

The ideas described above may help us to understand changes in the structure and intensity of a tropical cyclone during the transformation stage of ET. If the cyclone moves over cooler water during this stage, the surface entropy flux will decrease and boundary layer θ_e as well as its radial gradient will rapidly decline. If convection continues, the radial gradient of virtual temperature in the core throughout the troposphere will decline as well and the cyclone will weaken. If the cyclone moves over land during transition, the surface entropy flux will again decrease, but the relationship between the tropospheric virtual temperature and boundary layer θ_e hypothesized by Emanuel may not remain valid, except possibly in the eyewall.

The crucial role played by latent heat release in explosive extratropical cyclogenesis has been demonstrated in a number of studies (e.g., Reed and Albright 1986; Kuo and Reed 1988; Kuo and Low-Nam 1990; Shutts 1990; Kuo et al. 1991b; Gyakum et al. 1995; Davis et al. 1996; Rausch and Smith 1996; Carrera et al. 1999). These studies emphasize the nonlinear interaction between adiabatic dynamics and diabatic processes. Strong latent heat release occurs predominantly in slantwise ascent in the vicinity of the warm front in the presence of small symmetric stability. This results in enhanced ascent and descent, a sharp increase in the rate of surface frontogenesis, and an increase in low-level PV (Thorpe and Emanuel 1985; Emanuel et al. 1987). The magnitude of the frontogenesis is increased by both direct heating on the warm side of the front and by the diabatic enhancement of the ageostrophic frontal circulation. The increased frontogenesis leads to a strengthening of the adiabatic frontal circulation and additional latent heat release, so that there is a positive nonlinear feedback on the intensity of the extratropical system.

f. Underlying surface

The interaction between a tropical cyclone and the underlying surface is determined by the exchange of momentum, heat, and mass. Considerable effort has been devoted to studying the surface–atmosphere interaction for both tropical and extratropical cyclones, but the implications for ET have received little attention. In this section we describe the role of surface fluxes of momentum, heat, and moisture in both tropical and extratropical cyclones with emphasis on the relevance for ET. We discuss the role of surface fluxes in two cases of Atlantic ET and describe some of the uncertainties in our knowledge of the impact of the underlying surface.

Surface momentum fluxes act to decrease the intensity of a tropical cyclone as follows. Above the boundary layer in a tropical cyclone there is an approximate balance between the inward-directed pressure gradient force and the outward-directed Coriolis and centrifugal forces (Willoughby 1990). In the boundary layer, surface friction reduces the tangential wind so that the centrifugal and Coriolis forces are smaller than the pressure gradient force. The resulting net inward force drives a secondary circulation with strong inflow in the boundary layer and weak outflow above the boundary layer. As rings of fluid expand outward above the boundary layer, conservation of angular momentum requires that they spin more slowly (see, e.g., Holton 1992, section 5.4). The details of the vortex spindown depend on the surface boundary condition and the thermal stratification above the boundary layer. For a turbulent boundary layer and neutral stratification the inverse tangential wind speed should increase linearly with time and the rate of spindown is constant with height (Eliassen 1971; Eliassen and Lystad 1977; Montgomery et al. 2001). If there is stable stratification above the boundary layer, as would be the case when a tropical cyclone moves over cooler water and loses its inner-core convection, then the rate of spindown is strongest just above the boundary layer (Holton 1965). In addition, the presence of supergradient flow in the upper boundary layer (Kepert 2001; Kepert and Wang 2001) influences the outflow above the boundary layer. In the case when a tropical cyclone moves over cooler water at the beginning of ET, the above discussion suggests that the cyclone would spin down in the lower troposphere. The interaction of a moving tropical cyclone with the underlying surface leads to asymmetries in the boundary layer wind field (Shapiro 1983; Kepert 2001; Kepert and Wang 2001). The role of these asymmetries on vortex spindown during ET has not yet been investigated.

The direct effect of surface momentum fluxes on an extratropical cyclone is to weaken the low-level circulation and associated temperature anomalies (Farrell 1985, 1989b; Valdes and Hoskins 1988). Analytical and numerical models of idealized dry baroclinic waves show that a reduction in the strength of low-level anomalies leads to a reduction in the growth rate of unstable modes (e.g., Farrell 1985; Valdes and Hoskins 1998). Therefore, surface momentum fluxes might be expected to inhibit extratropical reintensification during ET.

The vital importance of the underlying ocean for the development and maintenance of tropical cyclones has been recognized for some time (Riehl 1950; Kleinschmidt 1951) and is underscored by studies that attempt to determine the upper bound on the intensity that a particular storm might achieve, the so-called maximum potential intensity (MPI; Emanuel 1988; Holland 1997). Tropical cyclones can be thought of as heat engines that extract energy in the form of sensible and latent heat from the warm ocean surface. Some of the energy is converted into kinetic and potential energy and the remainder is lost to space in the form of longwave electromagnetic radiation. Both Emanuel (1988) and Holland (1997) found that the MPI is a strongly increasing function of the sea surface temperature. A critical review of MPI theories is contained in the article by Camp and Montgomery (2001), who conclude that the Emanuel model (Emanuel 1995a and references therein; Bister and Emanuel 1998) is the closest to providing a useful estimate of maximum intensity. If ET takes place over the ocean, the sea surface temperature will generally decline as the storm moves poleward. Research is needed to determine whether MPI theory can be applied to a tropical cyclone in the transformation stage of ET.

The response of the ocean to a tropical cyclone can prevent a tropical cyclone from reaching its MPI. The oceanic response to the surface wind stress of a tropical cyclone results in the entrainment mixing of cooler thermocline water into the warmer oceanic mixed layer, leading to a decrease in SST (Ginis 1995 and references therein; Shay and Chang 1997; Shay et al. 1998; Jacob et al. 2000b). Coupled ocean–atmosphere numerical modeling studies have shown that this interaction has a negative impact on the intensity of a tropical cyclone (Sutyrin and Khain 1984; Khain and Ginis 1991; Bender et al. 1993; Hodur 1997; Emanuel 1999; Schade and Emanuel 1999; Bender and Ginis 2000). During ET, asymmetries in the wind field become more pronounced. Jacob et al. (2000a) showed that the ocean–atmosphere interaction in a tropical cyclone is sensitive to the asymmetric component of the wind field. Their numerical simulation of the ocean response to Hurricane Gilbert showed that the area-averaged fluxes of latent and sensible heat increased when the model was forced by the total wind rather than just the symmetric wind field (Fig. 17).

The impact of surface fluxes of latent and sensible heat on the development of midlatitude cyclones is not as clear as for tropical cyclones. Observational and modeling studies show that there are typically large upward fluxes of heat and moisture behind the cold front and ahead of the storm, with weak upward or even downward fluxes in the warm sector (e.g., Fleagle and Nuss 1985; Reed and Albright 1986; Nuss and Anthes 1987; Nuss 1989). This pattern of fluxes would tend to decrease the low-level temperature and moisture anomalies in a developing baroclinic wave, by warming and moistening the cold air and cooling and drying the warm air, and thus tend to decrease the intensity of an extratropical system (Danard and Ellenton 1980; Nuss and

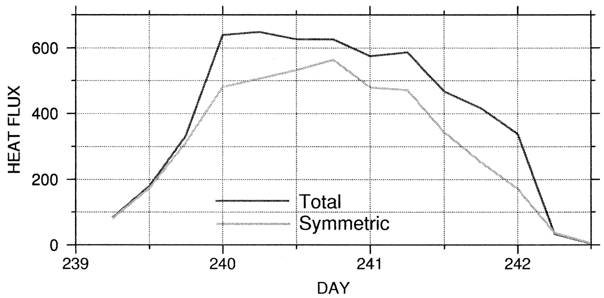

FIG. 17. Total heat fluxes from a simulation of the ocean response to the total and symmetric wind field of Hurricane Gilbert (1988). (Courtesy of D. Jacob.)

Anthes 1987; Chang et al. 1989; Nuss 1989). A number of studies of real cases of explosive cyclogenesis have reported that surface energy fluxes had little impact on the deepening rate during the period of explosive cyclogenesis (Kuo and Reed 1988; Kuo and Low-Nam 1990; Reed and Simmons 1991) but that the fluxes occurring in the 24 h preceding the period of rapid deepening, in particular to the northeast of the low center, made a significant contribution to the explosive deepening (Atlas 1987; Reed and Albright 1986; Nuss 1989; Nuss and Kamikawa 1990; Kuo et al. 1991a; Davis et al. 1996; Carrera et al. 1999). These results suggest that a crucial role of surface fluxes of heat and moisture is to precondition the environment into which the cyclone is moving to allow for strong latent heat release during the development of the cyclone by changing the stability and moisture content of the atmosphere. This preconditioning is consistent with the study of Davis and Emanuel (1988), who hypothesize that the surface fluxes act with convection to bring the atmosphere to a state of neutrality to moist convection. Bosart (1981) and Bosart and Lin (1984) discuss how heat and moisture fluxes acted to precondition the environment of the developing Presidents' Day storm (18–19 February 1979) by enhancing the amplitude of the lower-tropospheric coastal baroclinic zone and destabilizing the air mass ahead of the developing storm, thus contributing to the explosive development. In contrast to the previous studies, Chang et al. (1996) reported a positive impact of surface fluxes during the explosive intensification, which might be attributed to the development of a low-level hurricane-like vortex through air–sea interaction, as in Grønås (1995).

Thorncroft and Jones (2000) investigated the role of surface fluxes on the extratropical transitions of Hurricane Felix (1995), which filled steadily as an extratropical system, and Hurricane Iris (1995), which deepened explosively in the extratropics. The surface fluxes during the ET of Hurricane Iris illustrate the change from strong upward fluxes wrapped around the center of the tropical cyclone (Fig. 18a), to fluxes typical of an extratropical system (Fig. 18b). A similar pattern of surface fluxes to that seen in Fig. 18b was found in idealized numerical calculations of the transformation stage of ET (Ritchie and Elsberry 2001). Toward the end of the period of explosive deepening, the surface fluxes strengthened dramatically and, in addition to strong upward fluxes in the cold air to the west of the system, upward fluxes are seen wrapped around the center of the system. Thus in the inner core of the extratropical system the surface fluxes resemble those in a tropical cyclone (Fig. 18c). In addition, Thorncroft and Jones (2000) showed that at the time of Fig. 18c the inner core of the extratropical cyclone was characterized by a region of uniformly high equivalent potential temperature (θ_e) between the surface and the tropopause. The value of θ_e in this region was equal to the saturated equivalent potential temperature at the SST, leading Thorncroft and Jones (2000) to suggest that a similar air–sea interaction to that present in a tropical cyclone contributed to the explosive regeneration of the exhurricane. In contrast, Hurricane Felix moved over much colder water than Hurricane Iris early in its ET so that downward surface fluxes (Fig. 18d) toward the cool ocean acted to produce a stable boundary layer (and fog) and destroyed the warm core at low levels. The remnants of the tropical cyclone above the stable layer were then thermally ''decoupled'' from the ocean and decayed slowly until they moved into a warm frontal zone and were deformed horizontally. In the absence of large-scale horizontal or vertical shear the central warm core of strong wind, reduced midtropospheric static stability, and strong cyclonic vorticity may survive long after ET. Such cores can provide a nucleus for explosive redevelopment (via the modified Petterssen–Smebye type-B cyclogenesis discussed in section 3d) many days after a tropical cyclone remnant has ceased to be tracked (e.g., Pascoe et al. 1990).

Large uncertainty exists as to the magnitude of the air–sea exchange of momentum, heat, and mass at high wind speed. The key processes that influence air–sea fluxes are reviewed by Smith et al. (1996) and Geernaert (1999). The air–sea momentum exchange depends on the shear–flow interaction with surface waves, the wave breaking rates, and the direction of both the breaking wave events and the vertical wind shear. The air–sea heat and mass exchanges depend on evaporation from the surface, diffusive transfer across the laminar sublayer, evaporation of droplets emitted from bursting bubbles, evaporation of sea spray, turbulent transport efficiency, and surface cooling associated with gravitational settling of sea spray droplets that have been cooled by evaporation. Direct measurements of surface fluxes are available only for wind speeds of up to 25 m s^{-1}, and the uncertainty associated with the exchange coefficients for momentum, sensible heat, and latent heat at high wind speeds is large (Smith 1980; Geernaert et al. 1987; DeCosmo et al. 1996). Recent observational programs suggest that the direction of the surface wave field strongly influences the

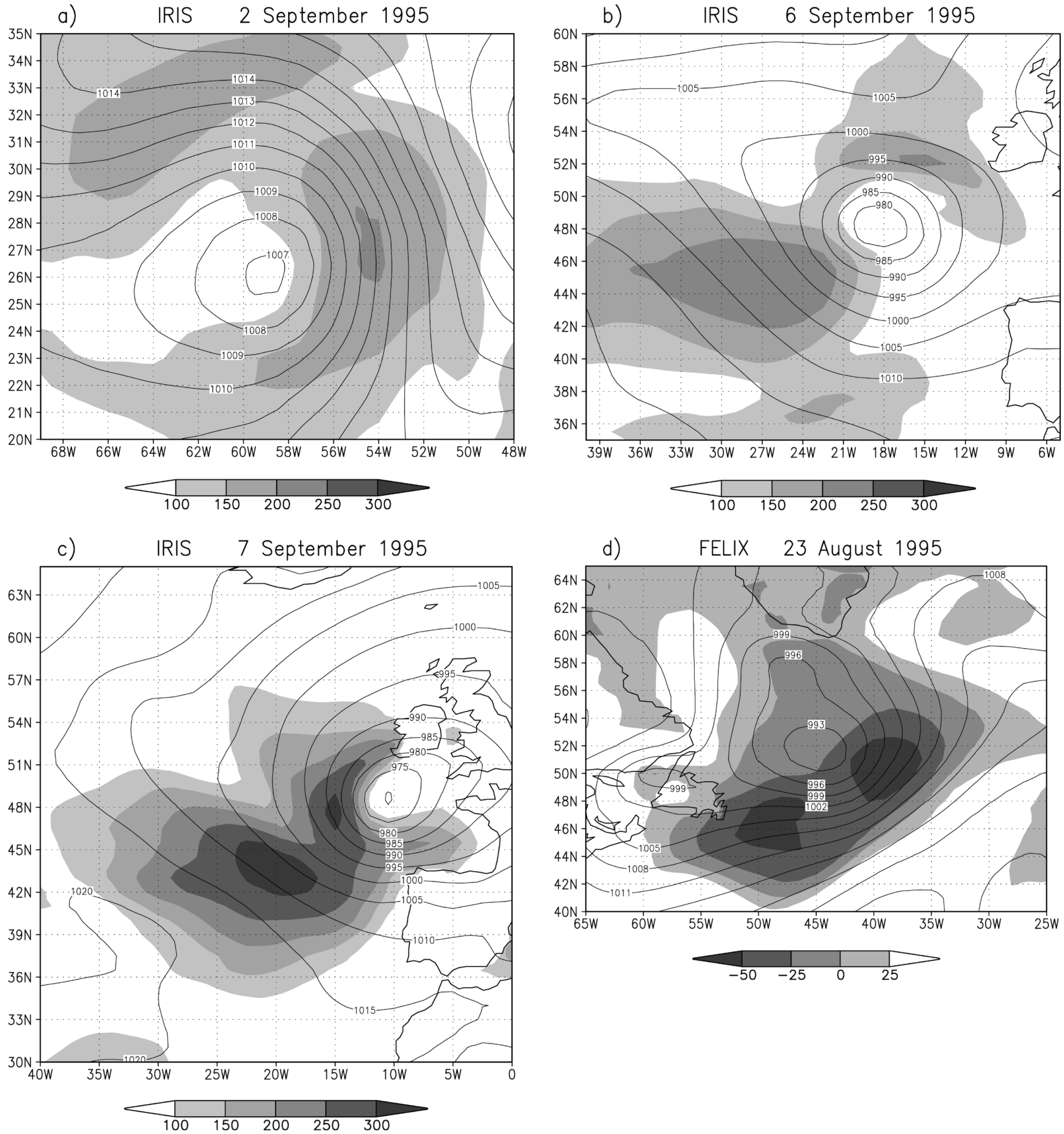

FIG. 18. Forecast surface latent heat fluxes in W m^{-2} (shaded) and analyzed mean sea level pressure in hPa (contours) at 0000 UTC on the dates shown for the extratropical transitions of (a)–(c) Hurricane Iris in 1995, where dark shading indicates strong upward fluxes, and (d) Hurricane Felix in 1995, where dark shading indicates strong downward fluxes. Data from ECMWF ERA-40 reanalysis (available online at http://www.ecmwf.int/research/era). Surface fluxes are a 6-h average centered on 0000 UTC from a forecast initialized at 1200 UTC on the previous day. [Adapted from Thorncroft and Jones (2000).]

direction of the surface wind stress vector (Geernaert 1988; Rieder et al. 1994).

The implications of the uncertainty in air–sea exchange coefficients for tropical and extratropical systems have received little attention. There has been some recent interest in the impacts of sea spray on tropical cyclones (Lighthill et al. 1994; Emanuel 1995a; Uang 1998; Bao et al. 2000; Andreas and Emanuel 2001; Wang et al. 2001), but there is at present no consensus as to whether sea spray causes a decrease or an increase in tropical cyclone intensity. Clearly more work is needed to clarify the influence on ET of sea spray and other factors that influence the interaction with the underlying surface.

g. Frontogenesis

The interaction between a tropical cyclone and the baroclinic midlatitude environment during ET can lead to the development of frontal regions. An indication of the structural changes that might occur during this interaction can be gained from kinematic analyses of the interaction between a tropical cyclone–like vortex and a straight baroclinic zone (Doswell 1984, 1985; Davies-Jones 1985; Keyser et al. 1988). The advection of the potential temperature field by the vortex results in the development of a frontal wave and of regions of enhanced temperature gradient reminiscent of cold and warm fronts in a typical extratropical system (Fig. 19). Keyser et al. (1988) analyzed the changes in the temperature field in terms of the vector frontogenesis function that consists of scalar and rotational contributions to frontogenesis. Scalar frontogenesis acts to change the magnitude of the potential temperature gradient while rotational frontogenesis acts to rotate the potential temperature gradient. Scalar frontogenesis may occur in association with divergence, horizontal deformation, and tilting whereas rotational frontogenesis may occur in association with relative vorticity, horizontal deformation, and tilting (Keyser et al. 1988; Schultz et al. 1998; Harr and Elsberry 2000).

Harr and Elsberry (2000) examined the scalar and rotational frontogenesis patterns associated with a variety of ET cases over the western North Pacific Ocean. They found a consistent overall pattern of scalar frontogenesis near the tropical cyclone, as illustrated here for the case of Typhoon David (1997). The maximum scalar frontogenesis (Fig. 20a) occurs north and east of the typhoon remnants. An area of scalar frontolysis occurs west of the center of the decaying tropical cyclone. The primary contribution to the scalar frontogenesis is due to horizontal deformation (Fig. 20b). To the north of the center of the decaying tropical cyclone, the axes of dilatation are nearly parallel to the isentropes, which is a deformation pattern consistent with frontogenesis in the warm frontal region when a cyclonic vortex is superimposed on a baroclinic zone (Keyser et al. 1988). Therefore, the deformation field associated with the movement of the tropical cyclone into the midlatitudes contributes to frontogenesis poleward and downstream of the tropical cyclone center, which promotes the development of a warm front. Warm frontogenesis poleward and to the east of the tropical cyclone center has been observed also in the southwest Pacific (Sinclair 2002). Upstream from the tropical cyclone center, the deformation field and the sinking motion associated with cold advection contribute to scalar frontolysis. The magnitude, areal extent, and structure of the frontogenesis in the overall pattern described above may differ between cases of ET, depending on the characteristic midlatitude circulation pattern into which the tropical cyclone is moving. In particular, the slope of the warm frontogenesis region depends on whether the cyclone is moving into a NE or NW pattern (Harr et al. 2000; Fig. 15).

The patterns of rotational frontogenesis are more variable between ET events. For the case of Typhoon David (representative of the NW pattern), there was a large region of positive rotational frontogenesis along the eastern boundary of the thermal trough with regions of negative rotational frontogenesis north and east of the typhoon center (Fig. 20c). The primary contribution to the positive rotational frontogenesis west of the decaying tropical cyclone center was horizontal deformation while the primary contribution to the negative rotational frontogenesis north and east of the center was anticyclonic vorticity downstream of the decaying tropical cyclone (Harr and Elsberry 2000). Positive rotational frontogenesis acts to rotate the gradient of potential temperature in a cyclonic direction, whereas negative rotational frontogenesis acts to rotate the potential temperature gradient in an anticyclonic direction (Keyser et al. 1988). Thus the pattern of rotational frontogenesis in Fig. 20c implies cyclonic rotation of the potential temperature gradient west of the typhoon center and anticyclonic rotation to the north and east, leading to an amplification of the thermal wave. The combination of rotational frontogenesis acting to amplify the thermal wave with scalar frontogenesis acting to enhance the horizontal temperature gradients indicates that reintensification as an extratropical cyclone is likely to occur (Harr and Elsberry 2000). In contrast, in the case of Typhoon Opal in 1997 (representative of the NE pattern) there was a large region of positive rotational frontogenesis to the north and east of the typhoon center (Fig. 20d). This implies a cyclonic rotation of the potential temperature gradient to the northeast of the tropical cyclone that would not lead to an amplification of the thermal wave. Thus the pattern of rotational frontogenesis could help to forecast whether or not reintensification will occur as an extratropical cyclone.

h. Precipitation

As discussed in section 2d, during an ET event the precipitation expands poleward of the center and is typically maximum to the left (right) of the track in the Northern (Southern) Hemisphere. The change in the structure of the precipitation field from the more symmetric distribution in a tropical cyclone to the asymmetric distribution during ET can be attributed to increasing synoptic-scale forcing of vertical motion associated with midlatitude features such as upper-level PV anomalies or baroclinic zones. This forcing can be expressed quantitatively in terms of the various forms of the quasigeostrophic omega equation (Hoskins et al. 1978 and references therein; Hoskins et al. 1985, see the appendix). DiMego and Bosart (1982a) diagnose the

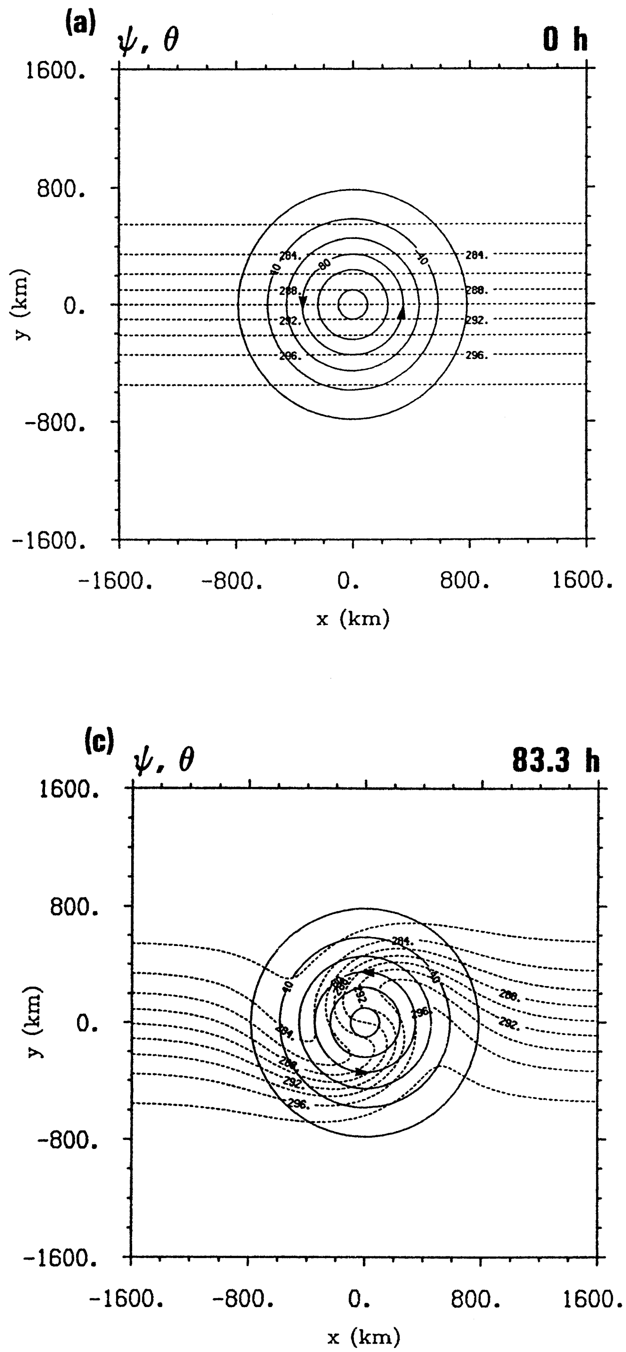

FIG. 19. The evolution of a baroclinic zone in a nondivergent steady-state vortex. Solid lines show the specified vortex streamfunction (contour interval is 20×10^4 m^2 s^{-1}); dashed lines show the potential temperature (contour interval is 2 K) at (a) 0, (b) 41.7, and (c) 83.3 h. [Taken from Keyser et al. (1988).]

different contributions to the vertical motion during the ET of Agnes (1972) and show how the forcing of vertical motion evolves from an almost symmetric forcing due to diabatic heating during the tropical phase to an asymmetric quasigeostrophic forcing during ET. The maximum ascent occurs where the region in which the vorticity advection increases with height overlaps the region of warm advection.

In the case of the ET of Hurricane Floyd (1999), Atallah and Bosart (2003) illustrate the increased quasigeostrophic (QG) forcing for ascent using both a Sutcliffe and a PV perspective framework. They showed that the heavy rainfall associated with the landfall and ET of Floyd shifted to along and to the west of a coastal front along which Floyd tracked. Use of a PV perspective enabled them to show that the coastal front was part of a deep baroclinic zone along the leading edge of the upstream PV anomaly. An intense 200-hPa jet ($>$75 m s^{-1}) and a strong ridge downstream of Floyd were manifestations of diabatically driven outflow from

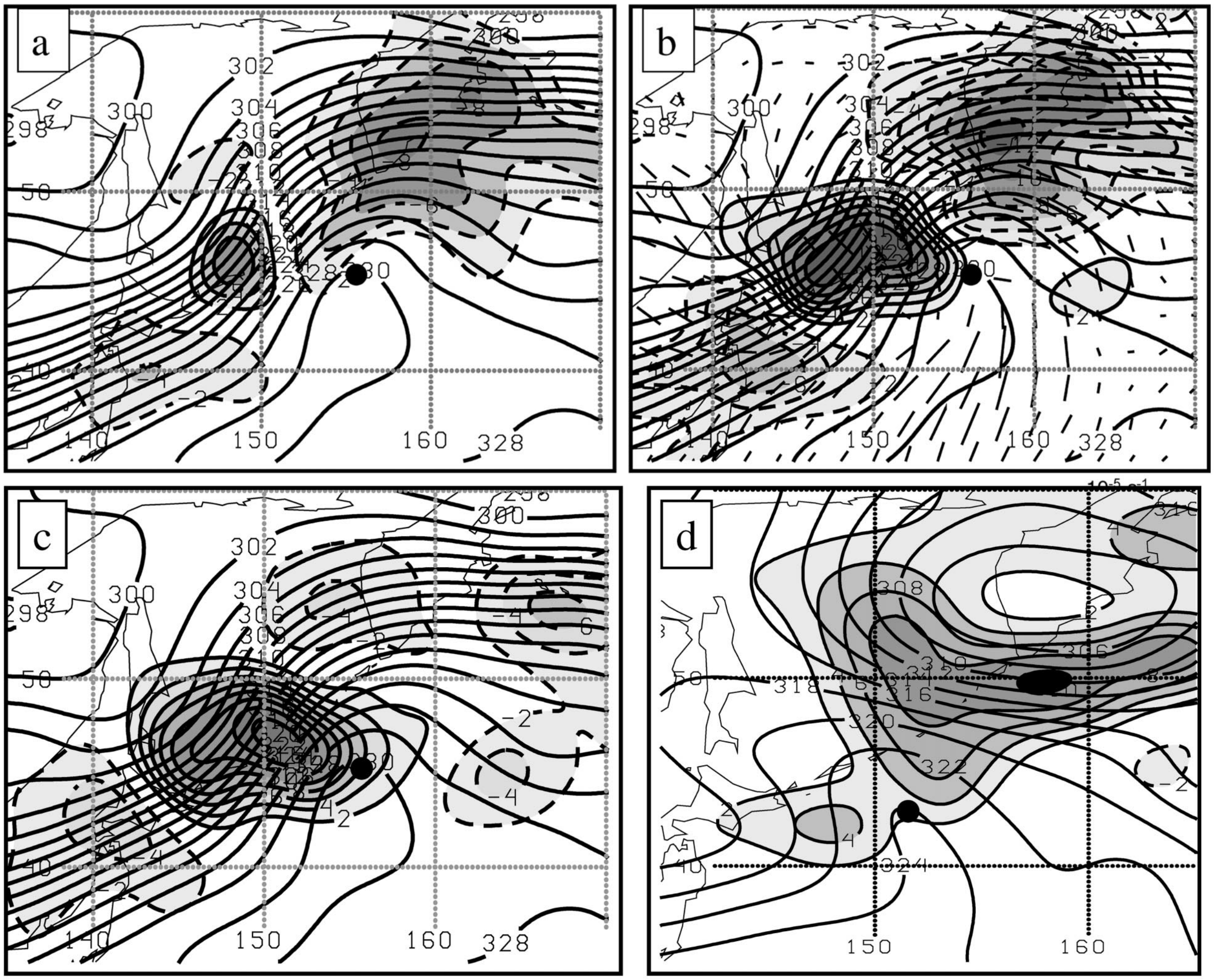

FIG. 20. Potential temperature at 500 hPa (contoured at 2-K intervals) and (a) shaded contours of scalar frontogenesis, (b) shaded contours of the contribution to scalar frontogenesis by horizontal deformation (negative values correspond to frontogenesis; positive values frontolysis), (c) and (d) shaded contours of rotational frontogenesis (positive rotational frontogenesis acts to rotate the gradient of potential temperature in a cyclonic direction; negative rotational frontogenesis acts to rotate the potential temperature gradient in an anticyclonic direction). Shaded contours are at intervals of 2×10^{-10} K m^{-1} s^{-1} with negative contours dashed. Axes of dilatation (see scale at lower right) are included in (b). The black dot marks the location of the low-level center of the typhoon. (a)–(c) For Typhoon David at 0000 UTC 20 Sep 1997 and (d) for Typhoon Opal at 1200 UTC 21 Jun 1997. [Adapted from Harr and Elsberry (2000).]

Floyd. The intensities of the 200-hPa ridge and jet downstream of Floyd were significantly underestimated in the National Centers for Atmospheric Prediction Eta and Aviation (AVN) Models, likely indicative of the inability of the models to simulate properly the bulk upscale effects of organized deep cumulus convection (see Fig. 15 of Atallah and Bosart 2003). The misrepresentation of the downstream ridging and associated jet in the numerical forecast can lead to inaccuracies in the quasigeostrophic forcing of vertical motion and thus in the precipitation forecast. A similar error pattern was noted in the case of the 12–14 March 1993 ''superstorm'' over eastern North America when massive convection over the Gulf of Mexico resulted in significant 200-hPa downstream ridging and jet development that was associated with widespread PV nonconservation on the dynamic tropopause (Dickinson et al. 1997).

Synoptic-scale forcing of ascent occurs below the right- (left-) entrance and left- (right-) exit regions of an upper-level jet in the Northern (Southern) Hemisphere. The movement of a tropical cyclone into such a region may be associated with enhanced precipitation (Sinclair 1993a, 2002; Atallah and Bosart 2003). Atallah and Bosart (2003) discuss how the diabatic modification of the upper-level PV discussed in section 3d can enable a tropical cyclone to remain beneath an equatorward jet entrance region. Sinclair (1993a) found that a concentrated region of ascent associated with heavy precipitation over northern New Zealand (Figs. 21a,b) during Tropical Cyclone Bola (1988) occurred between a dif-

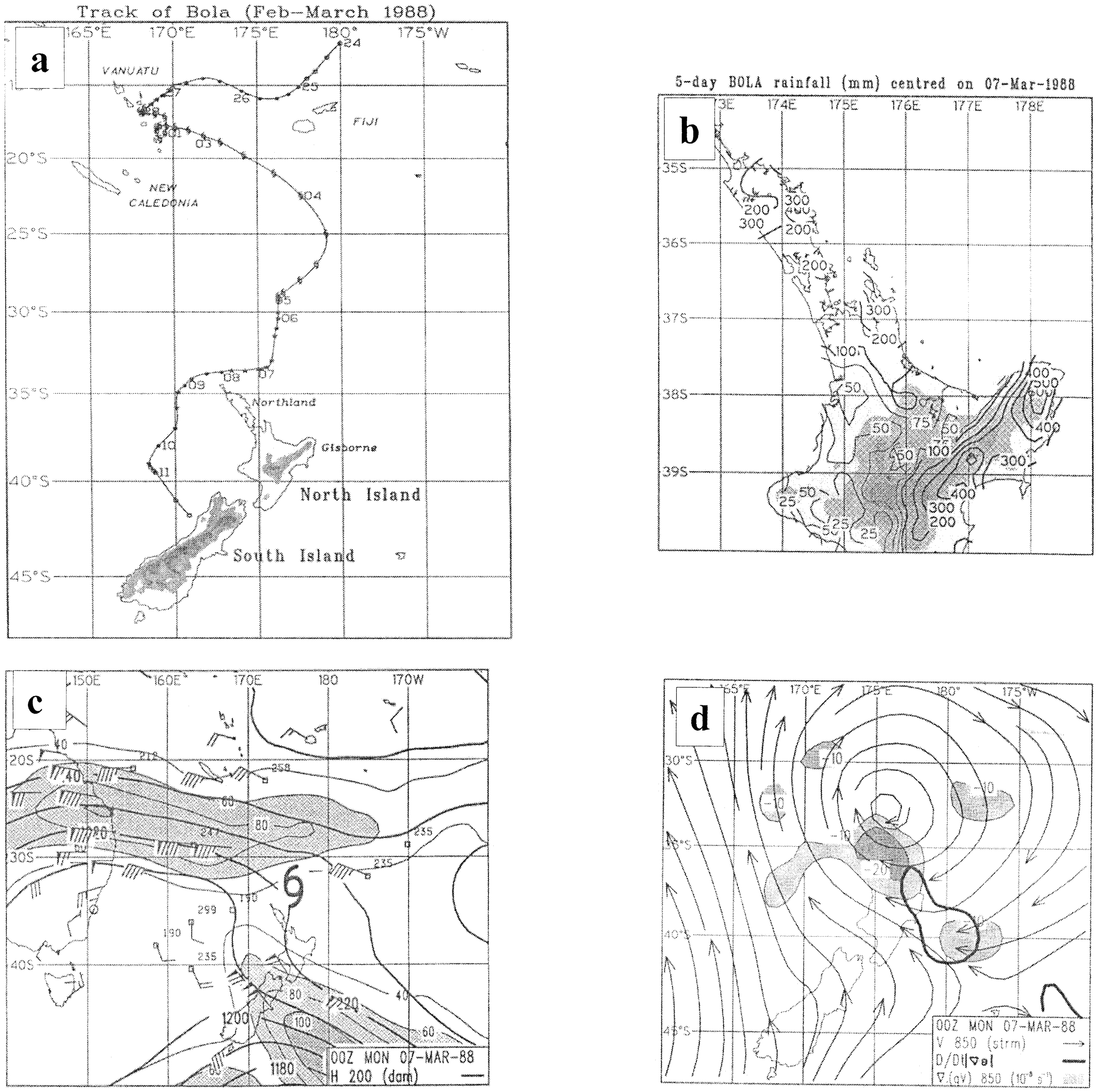

FIG. 21. (a) Track of Tropical Cyclone Bola (1988) based on 6-h positions with day numbers listed at the 0000 UTC position. (b) Rainfall (mm) over the North Island of New Zealand associated with the passage of Bola. (c) Isotachs (shaded with a contour interval of 20 kt) and heights (dynamical meters) at 200 h Pa at 0000 UTC 7 Mar 1988. (d) Streamlines at 850 hPa, moisture convergence (shaded in units of 10^{-8} s^{-1}), and region of 850-hPa total frontogenesis defined by a value of 3×10^{-10} K m^{-1} s^{-1}. [Adapted from Sinclair (1993).]

fluent upper-level jet exit region equatorward of Bola (Fig. 21c) and a confluent jet entrance region poleward of Bola, a location with strong quasigeostrophic forcing of ascent.

A majority of the precipitation associated with ET occurs poleward of the center of the decaying tropical cyclone. Harr and Elsberry (2000) identified this as a region of warm frontogenesis with the potential for increased forcing of vertical motion due to contributions from scalar and rotational frontogenesis. The structure of the warm frontogenesis region north and east of the tropical cyclone center is important since the majority of cloud and precipitation occurs over this region. Harr and Elsberry (2000) showed that the ascent and frontogenesis in the warm frontal region had a gentle upward slope. This suggests that warm, moist air in the southerly flow ahead of the tropical cyclone center ascends along the gently sloping warm front, allowing the region of precipitation to extend over a large area ahead of the tropical cyclone.

Since nearly all of the appreciable precipitation associated with ET occurs over the region of warm frontogenesis, it is important to consider contributions to the organization of precipitation over this region. One aspect to be investigated is the release of conditional symmetric instability (CSI; Bennetts and Hoskins 1979; Schultz and Schumacher 1999 and references therein) through the displacement of saturated air parcels along a slanted path (Emanuel 1983). The observation that the largest rain rates during ET tend to occur in bands that are oriented parallel to the thermal wind suggest that CSI may be an important factor during ET. In addition, a number of studies have shown that the combination of frontogenetical forcing and small symmetric stability leads to enhanced vertical motion that in turn acts to strengthen the frontogenesis (Emanuel 1985; Thorpe and Emanuel 1985; Sanders and Bosart 1985; Sanders 1986b).

Orography can lead to significant enhancement of precipitation during ET. For example, during the ET of Cyclone Bola a building anticyclone south of Bola caused the remnants of the tropical cyclone to turn west and contributed to strong onshore flow over the North Island of New Zealand (Fig. 21d). The topography of the island acted to concentrate the heavy rainfall along the eastern shore (Fig. 21b).

i. Energetics

A number of studies have calculated the kinetic energy budget during an ET event (Palmén 1958; Kornegay and Vincent 1976; Chien and Smith 1977; Vincent and Schlatter 1979; Harr et al. 2000). Palmén (1958) compared the ET of Hurricane Hazel with a typical extratropical cyclone in terms of their sources and sinks of energy. He found that an extraordinary amount of kinetic energy was exported to the midlatitude westerlies from the region of the decaying tropical cyclone, which led him to estimate that only two to three disturbances such as Hazel would provide the entire Northern Hemisphere north of 30°N with the kinetic energy sufficient to maintain the circulation against frictional dissipation.

An example of the kinetic energy budget for an ET event in which significant reintensification as an extratropical cyclone occurred is given here for Typhoon David (1997). In the analysis of Harr et al. (2000), the change in kinetic energy per unit area over an open region centered on the decaying Typhoon David was partitioned into a collection of sources and sinks. The primary components of the kinetic energy budget were defined as a generation (destruction) of kinetic energy (GKE) associated with cross-isobaric flow to lower (higher) pressure and the horizontal flux convergence (HFC) of kinetic energy from the surrounding region into the volume. A vertical flux term, a term related to the changes in mass of the volume, and a dissipation term were also included in the analysis. The vertical distribution of the primary terms in the kinetic energy budget throughout the ET of David (Fig. 22) indicates that the largest values of all terms occurred in the upper troposphere. The total kinetic energy (Fig. 22a) increased as the decaying tropical cyclone reintensified as an extratropical cyclone from 0000 UTC 19 September to 0000 UTC 20 September. Eventually, the time rate of change of kinetic energy became negative (Fig. 22b) when reintensification slowed. The sink of kinetic energy due to the HFC (Fig. 22c) may be attributed to the outflow from the decaying tropical cyclone (not shown) that exited northeast of the circulation center and appeared to contribute to increased southwesterly winds downstream from David. At the end of the reintensification of David as an extratropical cyclone (1200 UTC 20 September), the HFC term became small when the extratropical cyclone became aligned in the vertical (Harr and Elsberry 2000). The horizontal export of energy from the domain was partially offset by the generation of energy (Fig. 22d) associated with upper-level winds crossing from higher to lower heights (not shown). There was a maximum in GKE at lower levels also, which was partially offset by the residual dissipation term (not shown).

The horizontal export of kinetic energy from the region around the tropical cyclone during ET may be compensated by the generation of kinetic energy associated with both baroclinic and barotropic conversions. The analyses of Palmén (1958) and Harr et al. (2000) indicate that the coupling between the decaying tropical cyclone and the midlatitude circulation contributes to a conversion of available potential energy into kinetic energy via solenoidal circulations in which cold air sinks in the thermal trough upstream of the decaying tropical cyclone and warm air rises in the downstream thermal ridge. These are consistent with the role of frontogenesis in the amplification of the thermal wave during ET discussed in section 3g.

The analysis of the kinetic energy budget during ET suggests that there is a sensitive dependence of the budget on how the decaying tropical cyclone and midlatitude circulation interact. Variability in the tropical cyclone size and intensity may affect factors that have a significant impact on the kinetic energy budget, such as the remaining outflow from the decaying tropical cyclone or the strength of the solenoidal circulations.

4. Future directions

Extratropical transition is a complex evolutionary process that involves interactions over a variety of horizontal and vertical scales. Although much is yet to be learned about the thermodynamic and dynamic characteristics of a mature tropical cyclone, viable research programs are in place to examine these issues, as well as to examine the processes responsible for extratropical cyclogenesis. In contrast, the transition from a tropical

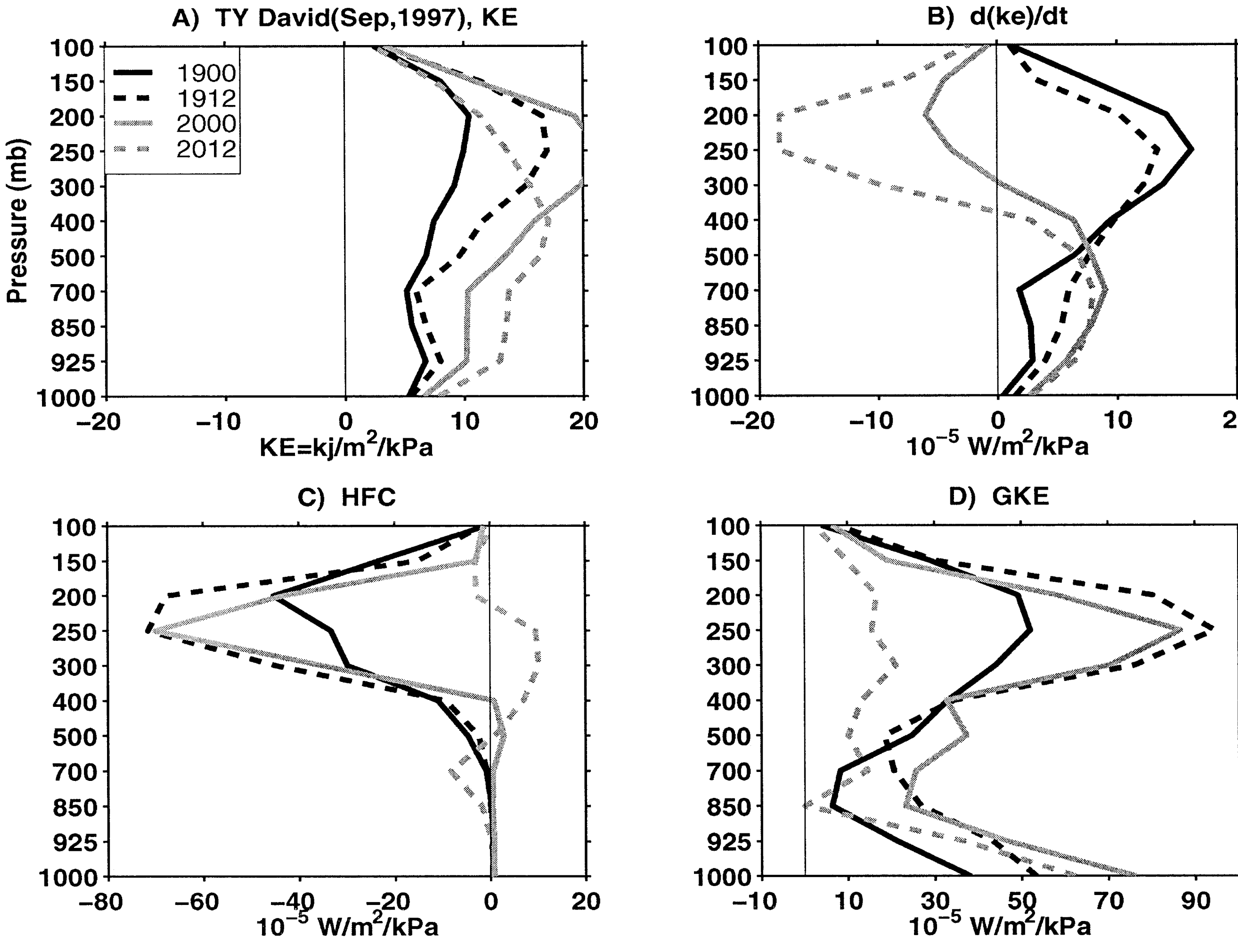

FIG. 22. Vertical distribution of the terms in the kinetic energy budget of Harr et al. (2000) for Typhoon David at the UTC times indicated in (a). (a) Total kinetic energy, (b) time rate of change of kinetic energy, (c) the horizontal flux convergence of kinetic energy from the surrounding region, and (d) generation (destruction) of kinetic energy associated with cross-isobaric flow to lower (higher) pressure. [Adapted from Harr et al. (2000).]

cyclone into an extratropical cyclone is poorly understood and incompletely researched.

A universal definition of ET does not exist. The evolutionary nature of ET and the time interval between successive observations and synoptic analyses makes it difficult, if not impossible, to specify a precise time at which a tropical cyclone has become extratropical. From an operational viewpoint, a definition that is based on tools available to a forecaster would be desirable. Possible frameworks for a definition are the two-stage description of ET (section 3a, Klein et al. 2000), the phase space diagram of Hart (2003), and the Sutcliffe approach of Darr (2002a–c).

Because the societal impacts of an ET event are due to specific physical conditions (e.g., precipitation coverage and amount, wind speeds, and wave heights) research needs to be focused on improved understanding of the evolution and prediction of these impact variables throughout the ET process. As discussed in section 2, the impact variables change in magnitude and distribution when a tropical cyclone moves poleward. Why and at what rate these changes occur remain important unanswered questions.

Many outstanding questions concern the changes in structure and intensity of a tropical cyclone during ET. Significant gaps exist in our understanding of tropical cyclone intensity change, both with regard to rapid intensification and to decay. The latter is expected to play a particularly important role in ET. Additional questions are related to the roles of complex terrain; convection; surface roughness; the air–sea exchange of heat, momentum, and mass; and the interaction with the ocean. Very little is known about the boundary layer environment at high wind speeds, and it is obvious that boundary layer processes will have an important impact on the evolution of precipitation, wind, and wave fields during ET.

Further research is needed to quantify whether changes in the nature of the tropical cyclone during the transformation stage influence the subsequent development during the extratropical stage. Klein et al. (2000) examined statistical relationships between the minimum central sea level pressure associated with the resulting extratropical cyclone and tropical cyclone characteristics such as intensity and size. Although their analysis was based on a limited sample of 33 ET cases over the western North Pacific, they could not identify a statistically significant relationship between the original tropical cyclone characteristics and the minimum central pressure of the extratropical system. However, their study considered only tropical cyclone characteristics before the start of the transformation stage, whereas tropical cyclone structure and intensity at the end of the transformation stage may have a larger impact on the extratropical stage. In addition, their study did not consider whether or not the midlatitude environment was favorable for extratropical development.

As a tropical cyclone moves poleward, the baroclinity of its environment increases and interactions with other synoptic features (e.g., fronts, upper-level troughs, mature extratropical systems) become more probable. The interaction between systems of tropical and extratropical origin and its influence on the primary variables that determine the impact of an ET event need to be investigated. A key question here is predicting the outcome when a midlatitude trough interacts with a tropical cyclone. Sometimes the vertical wind shear associated with the trough causes the cyclone to dissipate in the Tropics. Sometimes the trough can actually invigorate the tropical cyclone in the Tropics (without ET occurring). Sometimes the trough can ‘‘capture’’ the tropical cyclone, so that ET occurs with or without reintensification.

Following the transformation stage, reintensification as an extratropical cyclone may occur. Several studies have identified periods of reintensification that meet the criterion of rapid extratropical cyclogenesis (Sanders and Gyakum 1980). During extratropical reintensification, the relative roles of the remaining tropical cyclone features and the midlatitude characteristics in defining the reintensification as an extratropical cyclone need to be investigated.

An improved understanding of ET can be obtained through idealized numerical modeling, for example, with simplified physical representations or simplified initial conditions. Such an approach allows for particular processes to be studied in isolation. Then by gradually increasing the complexity of the model or of the initial conditions insight can be gained into the feedbacks between various processes. In addition, the further use of PV thinking promises to be of great value in the diagnosis of ET in idealized models, full physics models, and observations.

Often, predictions of ET by numerical forecast models do not accurately depict the characteristics of ET and the subsequent evolution of the resulting extratropical cyclone. Outstanding questions concern the representation of various physical processes that are likely to play an extremely important role during ET and are typically parameterized in numerical models. Also, increased understanding of the impact of data assimilation and of the inclusion of synthetic observations on numerical prediction of ET is required. Finally, investigation is required to establish the limits of predictability and to develop ensemble prediction techniques for ET. Since ET involves many different atmospheric processes, the inherent predictability of an ET event might be less than that of a pure tropical or midlatitude cyclone.

With regard to improved operational forecasts and warnings, outstanding questions remain as to how best to assimilate research results so that a forecaster can apply them to a specific ET scenario. Research is required to provide forecasters with conceptual models of ET and better diagnostic tools that would enable them to provide more effective advisories and warnings.

In order to further both the understanding and forecasting of ET it is essential that we improve our knowledge of real ET events, both by making better use of existing observations and by exploiting new observational capabilities. Investigations into ways of making better use of existing observations should consider both the documentation, validation, and development of multichannel passive and active satellite products, especially Advanced Microwave Sounding Unit (AMSU), TRMM, SSM/I, and scatterometer data, and improved data assimilation, including enhanced use of existing observations and efforts to further incorporate special observations. Improved knowledge of the detailed structure of ET systems requires more in situ observations of the evolution of the decaying tropical cyclone remnants and the midlatitude environment during ET. Further in situ observations could be obtained as part of a field experiment with intensive observation periods. The participation of both forecasters and researchers is essential for the success of this type of field program.

In this paper we have reviewed the forecast challenges associated with ET and the current understanding of the processes involved in ET. It is clear that there are many aspects of forecasting and understanding ET in which progress needs to be made. Substantial progress will be possible only through enhanced collaboration between the basic research and operational forecast communities.

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