Scale Interactions during the Formation of Typhoon Irving

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ABSTRACT

The development of Typhoon Irving is investigated using a variety of data, including special research aircraft data from the Tropical Cyclone Motion (TCM-92) experiment, objective analyses, satellite data, and traditional surface and sounding data. The development process is treated as a dry-adiabatic vortex dynamics problem, and it is found that environmental and mesoscale dynamics mutually enhance each other in a cooperative interaction during cyclone formation. Synoptic-scale interactions result in the evolution of the hostile environment toward more favorable conditions for storm development. Mesoscale interactions with the low-level, large-scale circulations and with other midlevel, mesoscale features result in development of vorticity in the midlevels and enhancement of the low-level vorticity associated with the developing surface cyclone.

Multiple developments of mesoscale convective systems after the storm reaches tropical depression strength suggests both an increase in low-level confluence and a tendency toward recurrent development of associated mesoscale convective vortices. This is observed in both aircraft data and satellite imagery where subsequent interactions, including mergers with the low-level, tropical depression vortex, are observed. A contour dynamics experiment suggests that the movement of mesoscale convective systems in satellite imagery corresponds well to the movement of their associated midlevel vortices. Results from a simple baroclinic experiment show that the midlevel vortices affect the large-scale, low-level circulation in two ways: 1) initially, interactions between midlevel vortices produce a combined vortex of greater depth; 2) interaction between midlevel vortices and the low-level circulation produces a development downward of the midlevel vorticity. This strengthens the surface vortex and develops a more cohesive vortex that extends from the surface through the midtroposphere.

1. Introduction

An interesting and complex aspect of tropical meteorology is the determination of the mechanisms by which a tropical cyclone develops. Although many studies of factors influencing tropical cyclogenesis have been completed, a consolidated theory has not been developed. Major difficulties include the following: 1) a lack of data over the large ocean regions; 2) long timescales, during which the appropriate large-scale conditions are satisfied without cyclogenesis occurring; 3) variations in the large-scale conditions between ocean basins that appear to affect the mechanisms by which cyclogenesis eventuates.

It is well known (e.g., McWilliams 1984) that in an environment without external forcing, turbulent two-dimensional flow has a tendency to form finite-amplitude, coherent structures. A random distribution of vorticity will concentrate into isolated vortices due to an upscale merger process in which larger vorticity perturbations in the flow sweep up the surrounding vorticity. This provides a potential mechanism for the development and growth of mesoscale vortices and initial development of tropical cyclones in the atmosphere without need for thermodynamic forcing. Such vortices, especially those in the midlevels, have a vertical to horizontal scale ratio of 1:100 (e.g., Raymond and Jiang 1990) and can be considered essentially two-dimensional, although they may exist at different atmospheric levels.

The simplest cases of vortex interaction are those of binary vortices. Fujiwhara (1921, 1923, 1931) used a series of laboratory experiments in water to show that vortices of the same rotation placed in near proximity to each other tend to approach in a spiral orbit that has the same sense of rotation as the original vortices. More recent work with contour dynamics methods has filled out much of the parameter space for such vortex interaction (Ritchie and Holland 1993; Melander et al. 1988;

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Waugh 1992). Provided the vortices are sufficiently close, the end result is a merged vortex that is both larger and more intense than either of the original pair. Binary tropical cyclones are perhaps the best documented type of interacting vortex pair, and Lander and Holland (1993) have shown that their interaction closely follows that which would be expected from two-dimensional theory. Similar processes have been observed and modeled for mesoscale atmospheric vortices (Holland and Lander 1993) and for interactions between large monsoon lows, mesoscale vortices, and tropical cyclones (e.g., the review in Holland 1995).

Atmospheric forcing, especially in the form of latent heat release, results in preferential regions of coherent vortex development. Such a region is the western North Pacific, where a stable relationship between convection, an active monsoon trough, and tropical convergence zone (Holland 1995) results in a variety of scales of mesoscale convective organization. Several such cloud clusters containing mesoscale convective systems (MCS) are often in evidence in satellite imagery at a single time period (Ritchie 1995). These cloud clusters are a well-known precursor to tropical cyclogenesis (McBride and Zehr 1981), but their role is poorly understood.

Recent field experiments in the western North Pacific have concentrated on collecting high-resolution, quality data in regions where tropical cyclones form (Elsberry et al. 1992; Dunnavan et al. 1992). U.S. Air Force (USAF) weather reconnaissance aircraft (WC-130) were used to collect data at a much higher resolution than is currently available using the standard observing network. These observations indicate that mesoscale vortices developing within MCS may be commonly associated with the genesis of tropical cyclones (Ritchie 1995).

Until recently, such mesoscale vortices have been largely ignored in the literature on tropical cyclogenesis. Due to the sparsity of data, studies have concentrated on the resolvable large-scale, environmental influences associated with cyclogenesis (e.g., Gray 1968, 1975; McBride and Zehr 1981; Tuleya and Kurihara 1981; Tuleya 1991). We view cyclogenesis as arising from a cooperative interaction between the larger-scale environment and mesoscale vortex dynamics (Ritchie 1995; Holland 1995) and will be reporting elsewhere on the details of such interactions. This case study examines the scale interactions occurring in the development of Typhoon Irving during the Tropical Cyclone Motion (TCM-92) field experiment in the western North Pacific (Dunnavan et al. 1992). Irving provides an excellent example of mesoscale vortex interactions combined with a cooperative scale interaction in an initially unfavorable large-scale environment.

Our aim in this paper is to use the formation of Typhoon Irving to provide an example of the complex interactions between vortices of differing scales in a cooperative feedback system. Section 2 provides a description of the data sources and models used in this study. A review of mechanisms associated with the development of midlevel vortices within the stratiform regions of MCS is provided in section 3. Section 4 contains a description of the pre-Irving environment and our investigation of the mesoscale dynamics that occurred. Section 5 provides a summary and discussion of our results.

2. Data and method

a. Data

The sparsity of data over tropical oceans has made validation of hypotheses related to mesoscale aspects of tropical cyclones difficult. The TCM series of field experiments were designed to collect data suited to testing hypotheses concerning mesoscale influences on motion and genesis of tropical cyclones (e.g., Elsberry et al. 1992). Thus, the requirements were for high-resolution data at multiple levels in the atmosphere preferably also at several consecutive times.

The main observing platform for TCM-92 was the USAF WC-130 weather reconnaissance aircraft. Flight-level data was sensed eight times per second and averaged into 10-s data files. These data were then averaged again into 1-min measurements of winds, temperature, dewpoint, and geopotential heights. Thus, at an air speed of 510 km h^{-1}, the data were sampled every 7 m and finally averaged to every 8.5 km. During each mission, the vertical levels at which high-resolution, flight-level data were collected were maximized. In general these were limited to the 300-, 500-, and 700-hPa levels to allow adequate horizontal coverage as well. Omega dropwindsondes were used to provide soundings of winds, temperature, humidity, and geopotential heights below flight level.

Quality control of the data was achieved in flight via the U.S. Air Force Improved Reconnaissance Weather System (IRWS). Degradation of flight data occurred mainly when the aircraft was turning or during icing or extreme wetting of the instrumentation. Thus, any data collected during such events were discarded from the dataset. Manual observations by the aerial reconnaissance flight officer during each sonde launch provided a quality-control check on any suspect sondes. More information on data collection and quality control can be found in McKinley (1992).

Three missions were flown into the precursor system for Irving over a period of 9 days. The first, on 24 July, investigated a pulsing region of cloud associated with a weak, large-scale surface circulation (McKinley 1992). The second, on 29 July, once again flew the precursor system when investigating a possible interaction between two weakly organized regions of cloud. The third, on 1 August, was flown when the developing surface circulation had intensified into a tropical depression.

Surface analyses that form the majority of the large-scale material for the study utilize standard ship and island data that were provided by the Joint Typhoon Warning Center (JTWC), Guam. Data available after the time of forecast including any dropwindsonde data were added to the charts prior to manual reanalysis.

The objective analyses that form some of the large-scale material for the study are derived from the Australian Bureau of Meteorology Tropical Analysis Scheme (TAS). The scheme is a three-dimensional univariate optimal interpolation analysis system, operated on a 2.5° grid with eight vertical sigma levels from 0.95 to 0.1. It utilizes all available ship and land data, satellite cloud-drift wind and temperature data, and conventional upper-air observations available at the time of analysis. Since operational analyses are used, no additional special aircraft observations are included in this dataset. For this reason, use of the TAS analyses is limited to the vertical levels between 250 hPa and the surface where manual analyses are not available. Details of the analysis method are given in Davidson and McAvaney (1981).

Infrared and visible imagery from the Japanese Geostationary Meteorological Satellite (GMS) were retrieved from the Bureau of Meteorology McIDAS system to supplement the large-scale analyses. In this study the pixel values were enhanced to emphasize temperatures colder than 212 K (−61°C) for identification of mesoscale features.

Forecast guidance, supplementary data, and analyses were provided by JTWC, which hosted the operations center for the TCM field experiments.

b. Barotropic modeling

Vortex patches, consisting of a region of constant vorticity surrounded by a potential flow, form the simplest type of finite-scale vortex. A familiar example is the Rankine combined vortex (Depperman 1947), which has a circular region of constant vorticity. The vortices used in this study have the form

$$\begin{aligned} V_T &= cr, \qquad r < R \\ \zeta &= 2c \\ V_T &= cR^2r^{-1}, \qquad r \geq R \\ \zeta &= 0, \end{aligned} \tag{1}$$

where V_T is the azimuthal flow, ζ is the vorticity of the patch, R is the radius of maximum winds, and c is a scaling constant. These vortices are described in detail in Ritchie and Holland (1993) and provide a good first approximation to the continuous vorticity profiles associated with deep-layer means of tropical cyclones (Ritchie 1995).

Interactions between multiple vortices are investigated by use of the contour dynamics method of solution for inviscid, incompressible fluids in two dimensions, following Zabusky et al. (1979) and Overman and Zabusky (1982). In this method the constant-vorticity patches are represented by the contour that surrounds each patch, reducing the dimensionality to one. The contour is approximated using discrete nodes and the method computes the dynamic interactions between the nodes directly requiring no use of boundary conditions. The system is nondimensional for computation and is applicable to vortices of any scale. Typical scaling values for a tropical cyclone are

$$\begin{aligned} \zeta_{\mathrm{tc}} &= 10^{-3}\zeta_n \ \mathrm{s}^{-1} \\ R_{\mathrm{tc}} &= 100R_n \ \mathrm{km} \\ v_{\mathrm{tc}} &= \frac{\nu_n\zeta_{\mathrm{tc}}R_{\mathrm{tc}}}{R_n} \ \mathrm{m\ s}^{-1} \\ t_{\mathrm{tc}} &= \frac{\zeta_n}{\zeta_{\mathrm{tc}}} \ \mathrm{s}, \end{aligned} \tag{2}$$

where v is the azimuthal wind, t is time, ζ and R are defined in (1), and subscripts tc and n refer to tropical cyclone values and normalized vortex model values, respectively. More details are given in Overman and Zabusky (1982) and Ritchie and Holland (1993).

c. Baroclinic modeling

Due to weak horizontal temperature gradients, tropical regions are often considered approximately barotropic for synoptic-scale motions (Holton 1992), although substantial vertical wind shear may be present. At the mesoscale, considerable vertical structure exists within MCSs (e.g., Menard and Fritsch 1989; Ritchie 1995; Harr and Elsberry 1996). Thus, while the barotropic approximation provides a good first-order approximation, especially for the many vortices that have high horizontal aspect ratios, there are potentially important and complex baroclinic interactions, including those that occur between systems located at different levels in the atmosphere.

In order to investigate some of these effects, we make use of the baroclinic model described in Holland and Wang (1995). This model solves the nonlinear, primitive equations on an f plane valid at 15°N formulated with Cartesian coordinates in the horizontal and a σ coordinate in the vertical. The model contains five layers from $\sigma = 0$ to 1, where σ is the terrain-following coordinate

$$\sigma = \frac{p - P_t}{P_s - P_t} \tag{3}$$

and the vertical boundaries are at $P_t = 100$ hPa and the surface pressure P_s. Horizontal velocity and potential temperature are defined in the middle of each layer and vertical velocity at the interfaces. All variables are carried at each of the 135 × 135 grid points, with a uniform spacing of 20 km. Sponge layers are applied to the lateral boundaries (Wang and Li 1992), and the method of ''circling smoothing'' is applied at each time step for nine boundary cycles (Holland and Wang 1995).

The model integration uses the explicit time-split algorithm of Gadd (1978) in which the governing equations are divided into adjustment and horizontal advection. Time integration of the adjustment stage is by a trapezoidal forward–backward scheme (Gadd 1980). Spatial differencing is achieved using a centered-difference scheme with second-order precision in the horizontal and the method of Arakawa and Lamb (1977) in the vertical. Integration of the advection stage uses a fourth-order Lax–Wendroff scheme, with splitting in the x and y directions. Moist processes and diabatic forcing are not included since we are interested in the dry-adiabatic vortex dynamics in this study.

Initiation of the model is accomplished by first defining the vorticity structure. The related mass and wind fields are then derived by the nonlinear balance inversion described by Wang (1995).

3. The development of mesoscale vortices: Theory and observations

The atmosphere within an MCS can be viewed as a potential-vorticity-conserving system (Raymond and Jiang 1990). Potential vorticity (PV) contains the combined effects of thermal stratification and local rotation:

$$q \equiv (\zeta + f)\frac{\partial \theta}{\partial z}, \tag{4}$$

where q is potential vorticity, ζ is the vertical component of relative vorticity, f is the Coriolis parameter, and θ is the potential temperature (Hoskins et al. 1985). Note that this is the small Rossby number, large Richardson number approximation of PV.

Latent heat release warms the stratiform-cloud region of an MCS and evaporation of rain cools the region below the cloud. As a result, the potential temperature gradient near the cloud base is sharpened. Because the stratiform cloud region is nearly neutral to moist processes (Chen and Frank 1993), gravity waves cannot be sustained and the local deformation radius approaches zero. Thus, the mesoscale convergence and stretching that develops just above cloud base (Houze 1977) produces a positive potential vorticity anomaly with substantial relative circulation on the scale of the stratiform cloud (Raymond and Jiang 1990).

Similar processes have been argued for convection (Raymond and Jiang 1990). In this case, the sharpened potential temperature gradient occurs near the boundary layer and this is where the potential vorticity anomaly develops. Hertenstein and Schubert (1991) modeled the effects of convective versus stratiform heating on the PV field and showed that although convective heating is more intense, its effect is weaker due to its short-lived nature. In their model, the convection produced a weak, near-surface PV anomaly, whereas the stratiform heating produced a strong, midtropospheric PV anomaly. The associated flows were correspondingly weaker (stronger) for the convective (stratiform) case. We speculate that in tropical regions, the energy due to the convection is more likely to be dispersed as gravity waves. If the local deformation radius is reduced by other means, such as in a preexisting surface cyclone, then intense convection could produce significant low-level PV anomalies on the scale of the cumulus cloud. However, for the precyclogenesis environments that are of interest here, we consider that the stratiform environment is necessary to form well-organized mesoscale vortices.

Fritsch et al. (1994) have discussed the maintenance of some MCSs by reinitiation of convection arising from favorable types of vertical wind shear in the environment. They propose that low-level environmental flow that is faster than the propagation speed of the midlevel vortex will flow up the isentropes until convective instability is reached. In the midlatitudes, these parcels only reach their level of free convection if they realize the maximum possible ascent by flowing into the center of the cold column beneath the PV anomaly. Thus, reinitiation of convection is often observed to be at the center of such vortices (Fritsch and Kain 1993). However, in the Tropics the lifting condensation level is at lower heights and convection is often observed to develop on the edge of the midlevel vortices (e.g., Harr and Elsberry 1996; Ritchie et al. 1995).

If the mid- to upper-level flow is in the same direction as the low-level flow relative to the vortex motion, then the convection may act to reinforce and enhance the vorticity perturbation leading to a stable, long-lived system (Raymond and Jiang 1990; Fritsch and Kain 1993). If the mid- to upper-level environmental flow is in a different direction to the low-level flow, the stratiform region develops away from the vortex, setting up the conditions for the formation of new vortices in the near vicinity of the first.

Potential consequences of the interaction between midlevel vortices in close proximity are discussed in detail in Ritchie and Holland (1993) and Ritchie (1995). Such systems can undergo quite complex interactions consisting of mutual orbit, merger, and filamentation. In the right conditions, as outlined in previous paragraphs, interacting vortices may induce new convection that reinforces each vortex or leads to the development of new vortices. Some of these processes are described in sections 4c and 4d.

Meteorologists have yet to document the full life cycle of MCS-generated vortices. However, clues to their structure and intensity can be found in various case studies (e.g., Menard and Fritsch 1989; McKinley 1992; Harr and Elsberry 1996). McKinley (1992) found that a well-developed, 300–700-hPa vortex developed on the scale of 100–200 km within the stratiform region of a long-lived (12–18 h) MCS during TCM-92. Comparison with a less-organized, shorter-lived system revealed a smaller, weaker vortex of less vertical extent but with a similar thermodynamic structure. No information on the strength of these vortices is available. However, Menard and Fritsch (1989) and Harr and Elsberry (1996) found a maximum relative vorticity of about 3.0×10^{-5} and 9.0×10^{-5} s^{-1}, respectively, in the midlevels of the vortices they investigated. Note that the case Harr and Elsberry (1996) investigated was embedded within the circulation of a tropical cyclone and thus may be an overestimate of strength.

The decay of MCSs has not been documented. In fact, Lander and Holland (1993) document such vortices as traceable in satellite imagery for as long as 5 days in the western North Pacific. During this time, little convective activity was observed to be associated with the vortices. It is interesting that although these vortices developed within the region of high ambient vorticity associated with the monsoon trough, they retained their distinct identities for a long period of time. We surmise that although their size was so much smaller, their vorticity was strong enough that they were not simply absorbed into the large-scale pool of vorticity associated with the monsoon trough (Ritchie and Holland 1993). It is thus reasonable to investigate the direct dynamic impact of such mesoscale vortices on cases of tropical cyclone formation in which marked MCS development is observed. The following study of Irving's development will proceed with this in mind.

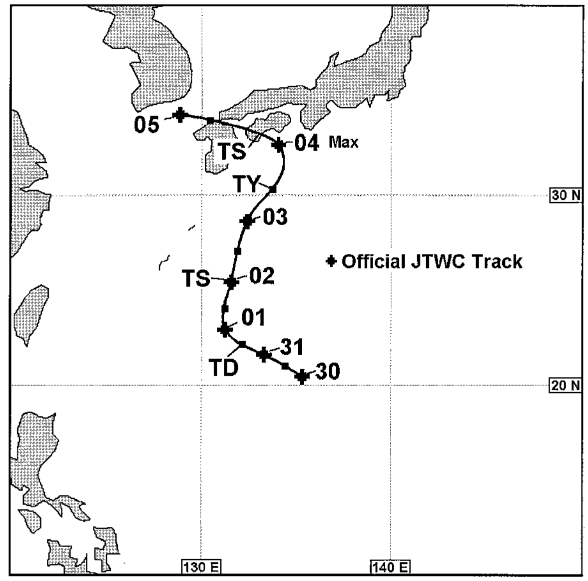

FIG. 1. The official JTWC track of Typhoon Irving. All times are UTC.

4. The development of Typhoon Irving

a. Synoptic conditions

The large-scale and climatological environments conducive for a tropical cyclone to develop include ocean temperatures of greater than 26°C, enhanced low-level cyclonic vorticity, midlevel moisture and conditional instability through a deep layer, and a region of disturbed weather with upper-level divergence and low-level convergence (Gray 1968, 1975). Once a mesoscale region of low pressure associated with the incipient cyclone has developed, the ocean can release additional energy to maintain an intensification cycle (Byers 1937; Kleinschmidt 1951; Emanuel 1986; Holland 1995). The initial development is complex and may involve interactions across several scales, varying from one case to another. For example, Holland (1995) has shown how an active monsoon trough, Rossby wave group and phase propagation, easterly waves, and even prior tropical cyclones can all interact to enhance the potential for tropical cyclone formation in the western North Pacific region.

Irving developed out of one of the many synoptic-scale surface cyclonic gyres propagating through the western North Pacific in late July 1992. Generally suppressed conditions existed throughout the region, which was dominated by a large upper-level anticyclone. Sporadic convection associated with one of the low-level gyres appeared to have been the trigger for the development of Irving into a tropical depression. Subsequent formation of MCSs within the tropical depression circulation resulted in slow development and intensification.

As shown in Fig. 1, Irving reached tropical storm strength on 2 August while moving on a slow northward track. It attained its maximum intensity on 4 August at a time when satellite imagery indicated an elliptical eye of 150-km diameter. Irving subsequently made a sharp turn to the west across southern Japan, rapidly weakened, and dissipated near Korea. Postanalysis indicates that Irving reached a minimum sea level pressure of 975 hPa and maximum surface winds of 40 m s^{-1} (ATCR 1992).

Irving developed in a relatively hostile environment by climatological standards. Anomalously high pressures of 1010–1012 hPa extended over the entire monsoon trough region east of the Philippines for all of July. Associated surface and gradient-level winds were weak at approximately 3–6 m s^{-1} (Fig. 2). There was no evidence in the objective analyses of the large-scale region of enhanced low-level cyclonic vorticity usually associated with the western North Pacific monsoon trough. However, a progression of weak, synoptic-scale, low-level circulations (LLCs) were observed to propagate across the region in the weeks prior to Irving's development (e.g., Figs. 2a,b). Development of a second circulation near one of these circulations resulted in a stronger, more compact low-level system. This gradually developed into Irving with a tropical storm intensity ($>$17 m s^{-1} sustained winds) on 2 August.

The upper-level flow was dominated by a large anticyclone over the region early in the development of

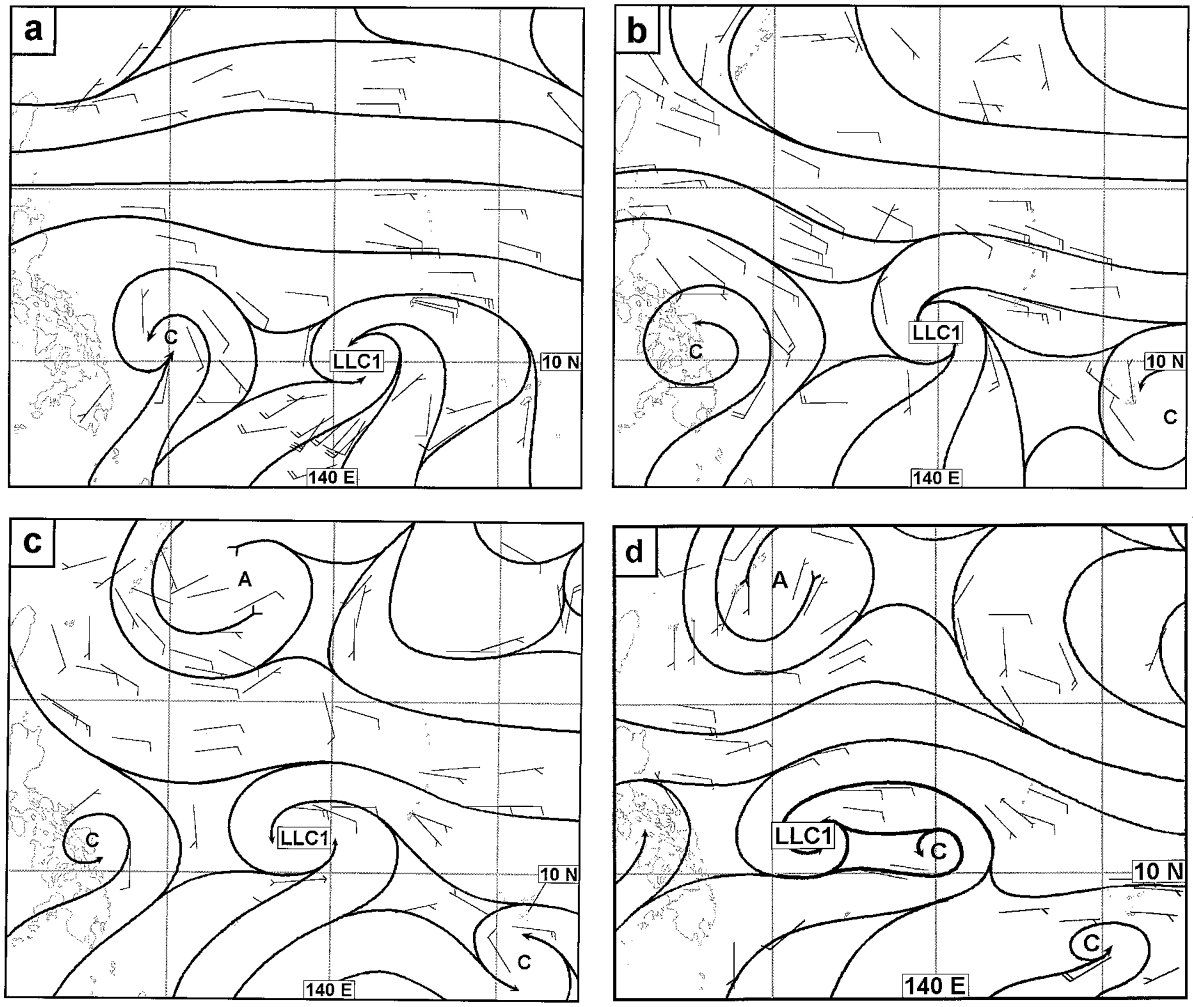

FIG. 2. Manual surface streamline analyses at 0000 UTC for the period 24–31 July 1992: (a) 24 July, (b) 25 July, (c) 26 July, (d) 27 July, (e) 28 July, (f) 29 July, (g) 30 July, and (h) 31 July. The low-level circulation initially located near 140°E is labeled LLC1. Latitude and longitude lines are plotted at 10° intervals. A full wind barb is 5 m s^{-1}.

Irving (Fig. 3a). Convection was suppressed in the domain poleward of 10°N and east of the Philippines. By 27 July (Fig. 3b), a tropical upper-tropospheric trough (TUTT) had moved into the region and developed strong convection within its circulation.

Interestingly, the low-level circulation that provided the background feature in which Typhoon Irving developed was analyzed one week before formation occurred. Although the circulation was initially at the eastern edge of the monsoon trough, an area generally favorable for the development of tropical cyclones (Briegel 1993; Holland 1995; Ritchie 1995), development did not occur immediately. The large-scale environment first became more favorable for the development of organized convection and MCSs before cyclone development occurred. It may be that in this ''slow-motion'' case of development it will be possible to examine the mesoscale dynamics of tropical cyclogenesis that ordinarily occur too rapidly for the standard observing network to capture.

b. Development of the pre-Irving depression

1) LLC DEVELOPMENT STAGE: 23–29 JULY

Two low-level circulations were evident in the analyses on 24 July (Fig. 2a). We designate the eastern circulation that we are interested in following as LLC1 (near 140°E). By 27 July, the western LLC had propagated westward out of the area of interest and LLC1 had moved westward to near 132°E.

Clusters of convection developed during 23 and 24 July (Fig. 4) near LLC1. The circulation associated with LLC1 was observed to extend up to 700 hPa in the TAS analyses, but no mesoscale details were resolved by the 2.5° latitude resolution employed. Aircraft data for a 10-h period centered on 0000 UTC 24 July verified that

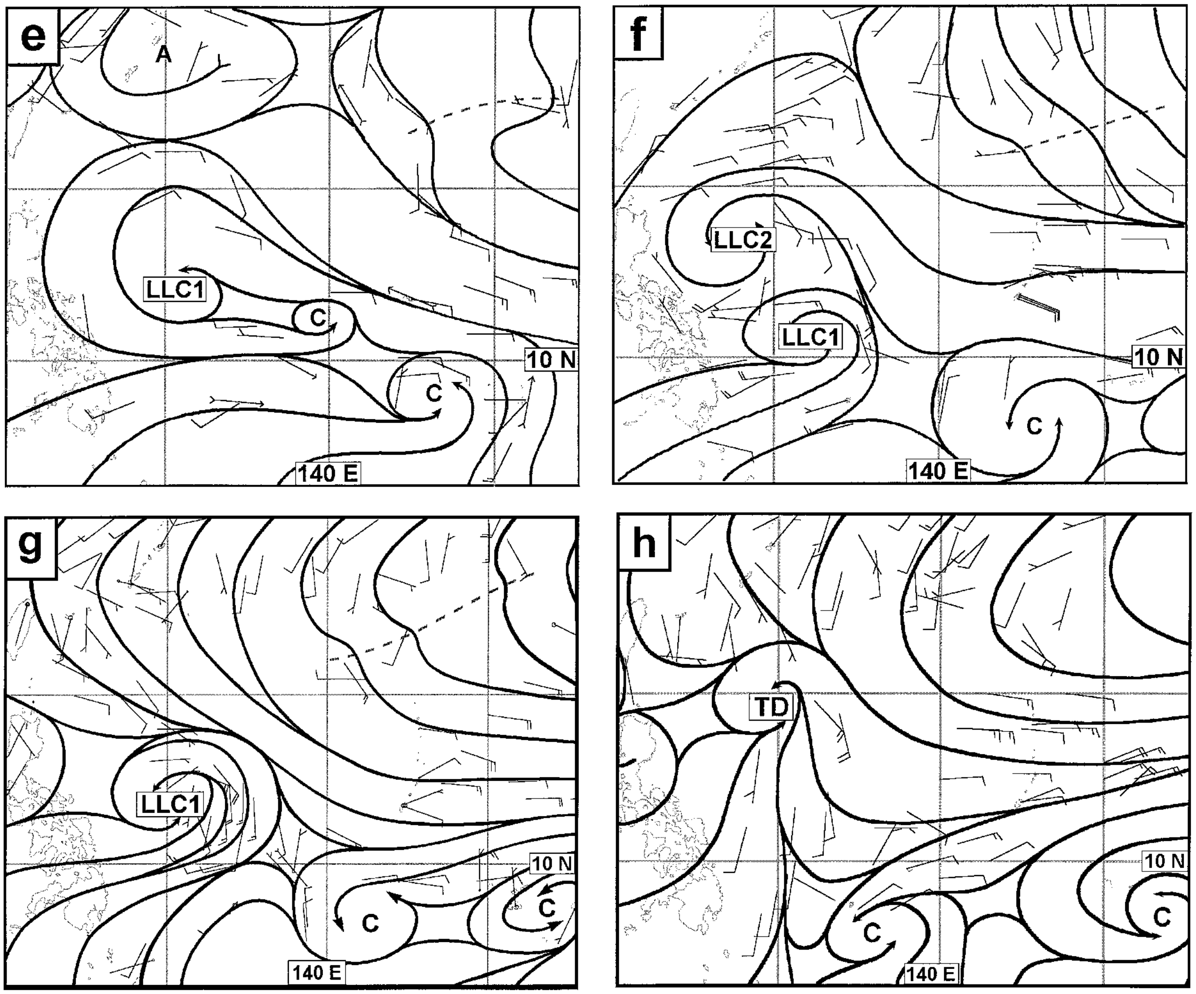

FIG. 2. (*Continued*)

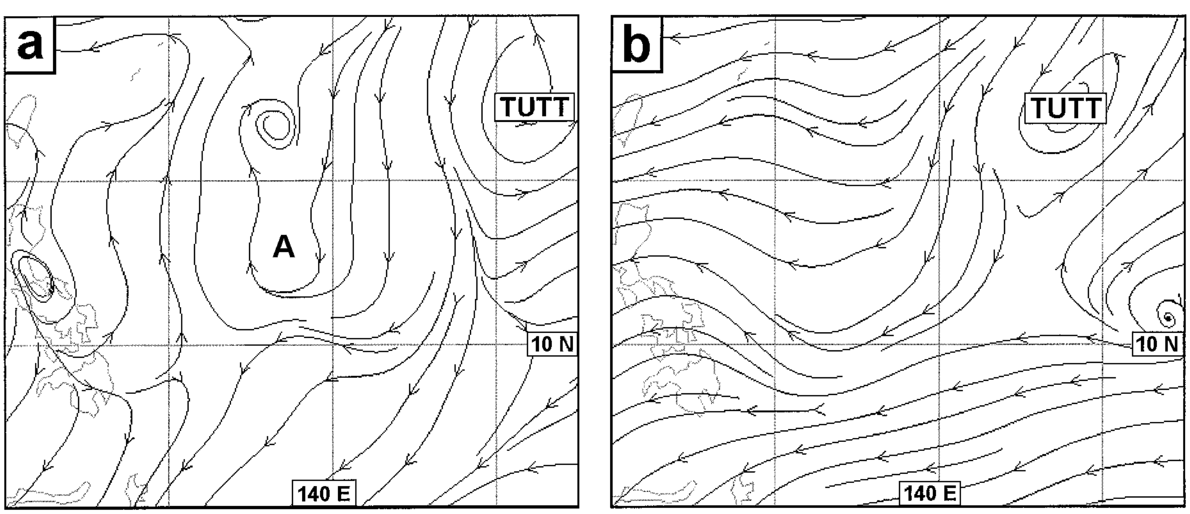

FIG. 3. Streamline analyses at 250 hPa for 0000 UTC from the Tropical Analysis System: (a) 25 July 1992 and (b) 27 July 1992.

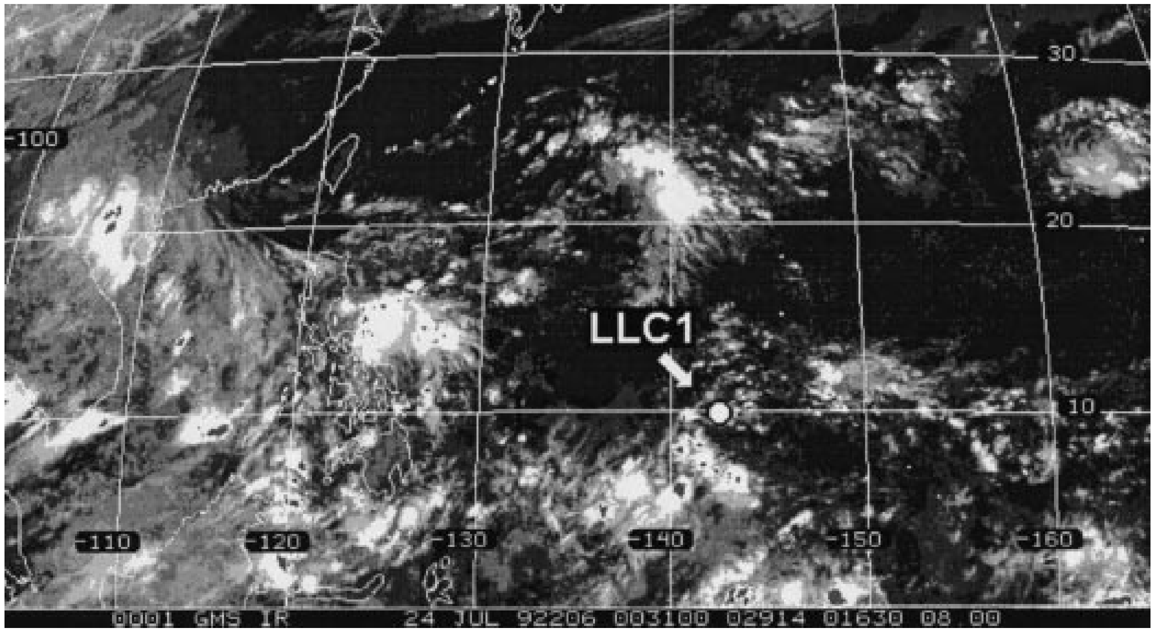

FIG. 4. Satellite imagery for 0000 UTC 24 July for the western North Pacific region. The circle indicates the position of LLC1.

the low-level circulation was about 700 km in diameter. This circulation extended from the gradient level to about 500 hPa (Figs. 5a,b), but with substantial shear and reduction in size. Other mesoscale vortices were found at 500 (Fig. 5a) and 300 hPa (not shown) in the aircraft data near the low-level circulation, with diameters of between 150 and 250 km. No mesoscale vortices were observed below 500 hPa, but this may be due to the coarser resolution of the dropwindsonde data. We suggest that it is unlikely that these mesoscale vortices arose from a reflection upward of the low-level vortex. Rather, the low-level circulation provided an environment for development of weakly organized MCSs (Fig. 4), which then developed mesoscale vortices by the processes discussed in section 3. Temperature and dewpoint analyses of the aircraft data by McKinley (1992) exhibit high moisture and a warming of 2°–3°C in the midtroposphere, supporting the MCS development theory. A surface cooling of a similar degree was also found by McKinley, with moist pockets of cool air from the sur-

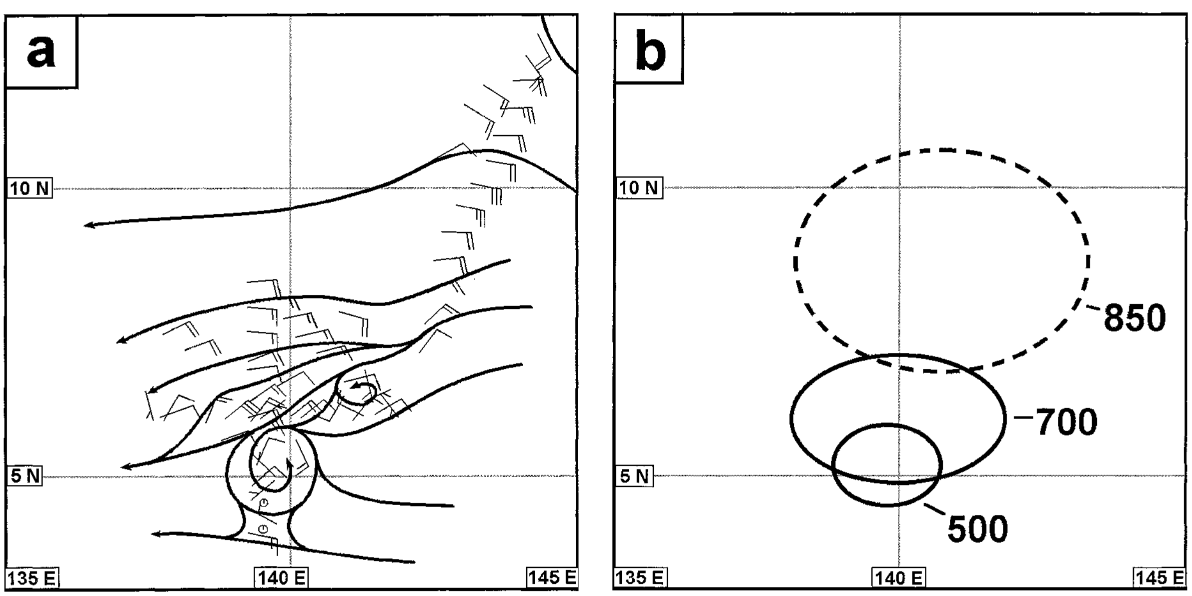

FIG. 5. (a) Streamline analysis of 500-hPa aircraft data for a 10-h period centered on 0000 UTC 24 July. (b) Relative sizes and positions of the 500-, 700-, and 850-hPa circulations as indicated in aircraft and surface data.

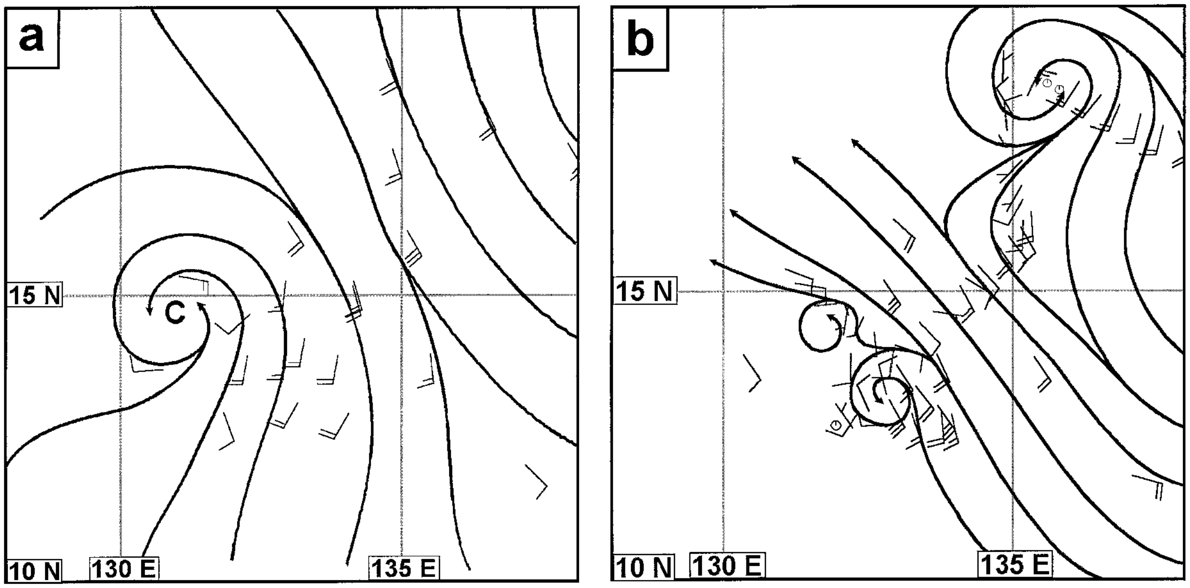

FIG. 6. Streamline analysis of aircraft data for a 10-h period centered near 0000 UTC 30 July: (a) 850 and (b) 500 hPa.

face to 700 hPa. Provided the vertical shear of wind is favorable, the cold pools enable uplift of moist surface air into the system, enhancing its ability to sustain convection as described in section 3. Development and decay of multiple MCSs continued through 26 July, then ceased briefly on 27 July.

A secondary maximum of convection near 13°N, 132°E during 26 July may have been responsible for the development of the second surface cyclone to the northwest of LLC1 (Fig. 2f). Although this is purely speculation at this time, the observations clearly show the development of cyclonic turning in the winds to the east and north of LLC1 in Figs. 2d–f, which by 29 July is a distinct second circulation labeled LLC2. However, by 0000 UTC 30 July, only one vortex could be observed near 14°N, 130°E (Fig. 2g). This new system contained winds of up to 12 m s^{-1} in a more compact region and was located at the centroid between the two LLCs. Thus it is possible that this vortex developed as the result of merger between LLC1 and LLC2. It is also possible that LLC2 simply propagated out of the region on this time frame. This seems unlikely, however, given that a translation speed of more than 12 m s^{-1} would be required. This would also leave no explanation for the increase in strength and northward movement of LLC1.

Aircraft data for the 9-h period centered on 0000 UTC 30 July verified that the systems had merged into a single vortex with an asymmetric maximum wind of about 11 m s^{-1} in the southeastern quadrant (Fig. 6a). This is consistent with the findings in Ritchie and Holland (1993) that the strongest winds initially develop in the vicinity of the shearing vortex.

Two midlevel vortices at 500 hPa were found by the aircraft within the merged LLC system (Fig. 6b). These vortices were small (<100 km) scale and no significant anomalies in air temperature or standard pressure height were associated with them. No major convective activity was visible in satellite imagery in their vicinity in the 24 h prior to the aircraft flight. The lack of temperature and moisture anomalies associated with these vortices also supports this and suggests that the vortices had existed for at least 48 h prior to the aircraft flight. It is possible that these vortices were the remnants of those first observed in the aircraft data from 24 July. We can speculate that the lack of a shearing environment enabled these vortices to exist as distinct entities throughout this period. The sporadic convective events that occurred between 25 and 28 July in the vicinity of LLC1 would have served to enhance these vortices via mechanisms discussed in Fritsch et al. (1994). However, lack of data denies us the opportunity to test the convective suppression–initiation roles of such vortices.

2) TUTT INTERACTION STAGE: 30–31 JULY

The large vortex in the northeast corner of Fig. 6b arose from downward penetration of a closed cell within the encroaching TUTT in Fig. 3b. The vortex was centered at 18.6°N, 135.6°E at 500 hPa, with maximum winds of 8 m s^{-1} at a radius of 40 km. Relative humidity at this level was close to 100% in the vicinity of the vortex, but there was little anomaly in the air temperature and standard pressure height. There was no observable circulation below 500 hPa (Fig. 6a).

An interaction between the LLC and this TUTT-generated, midlevel vortex resulted in the development of the tropical depression. The sequence of satellite imagery and circulation centers in Fig. 7 shows these two systems merging while generating a number of MCSs.

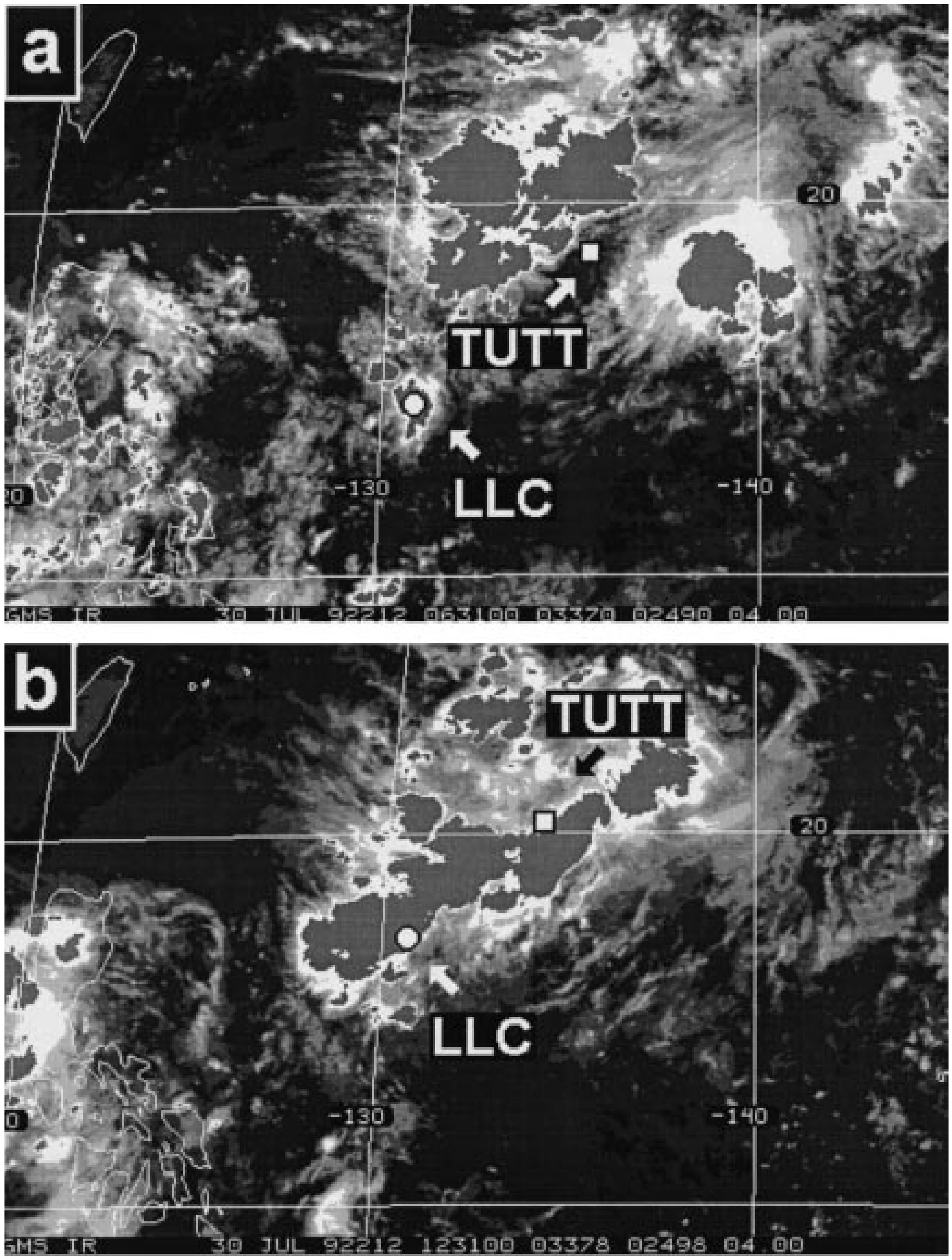

FIG. 7. Satellite imagery of the interaction between the low-level pre-Irving circulation (○) and the midlevel circulation associated with the TUTT (□): (a) 0600 UTC 30 July, (b) 1200 UTC 30 July, (c) 2100 UTC 30 July, and (d) 0000 UTC 31 July.

By 0000 UTC 31 July (Fig. 7d) a clearly defined, low-level circulation was evident in the satellite imagery. It is likely that the convection associated with the merger of the LLC and TUTT-generated systems helped transfer PV from the upper-level TUTT downward to increase the intensity of the surface vortex, but we have no way of confirming that this did indeed occur.

c. Development to tropical storm

The development of the storm from depression to storm strength was marked by growth and interactions of a number of MCSs (Figs. 8a–e). The track of each cluster (Fig. 9) indicates that they orbited the synoptic-scale cyclone center. The systems to the east merged toward the center while being sheared into a long filament shape. Those to the southwest were advected away from the synoptic center by the elongated region of northerlies in that region (Fig. 10). Our earlier discussion on reinitiation of convection in mesoscale convective vortices suggests that it is reasonable to assume that these MCSs were associated with such vortices. The aircraft reconnaissance centered on 0800 UTC 1 August

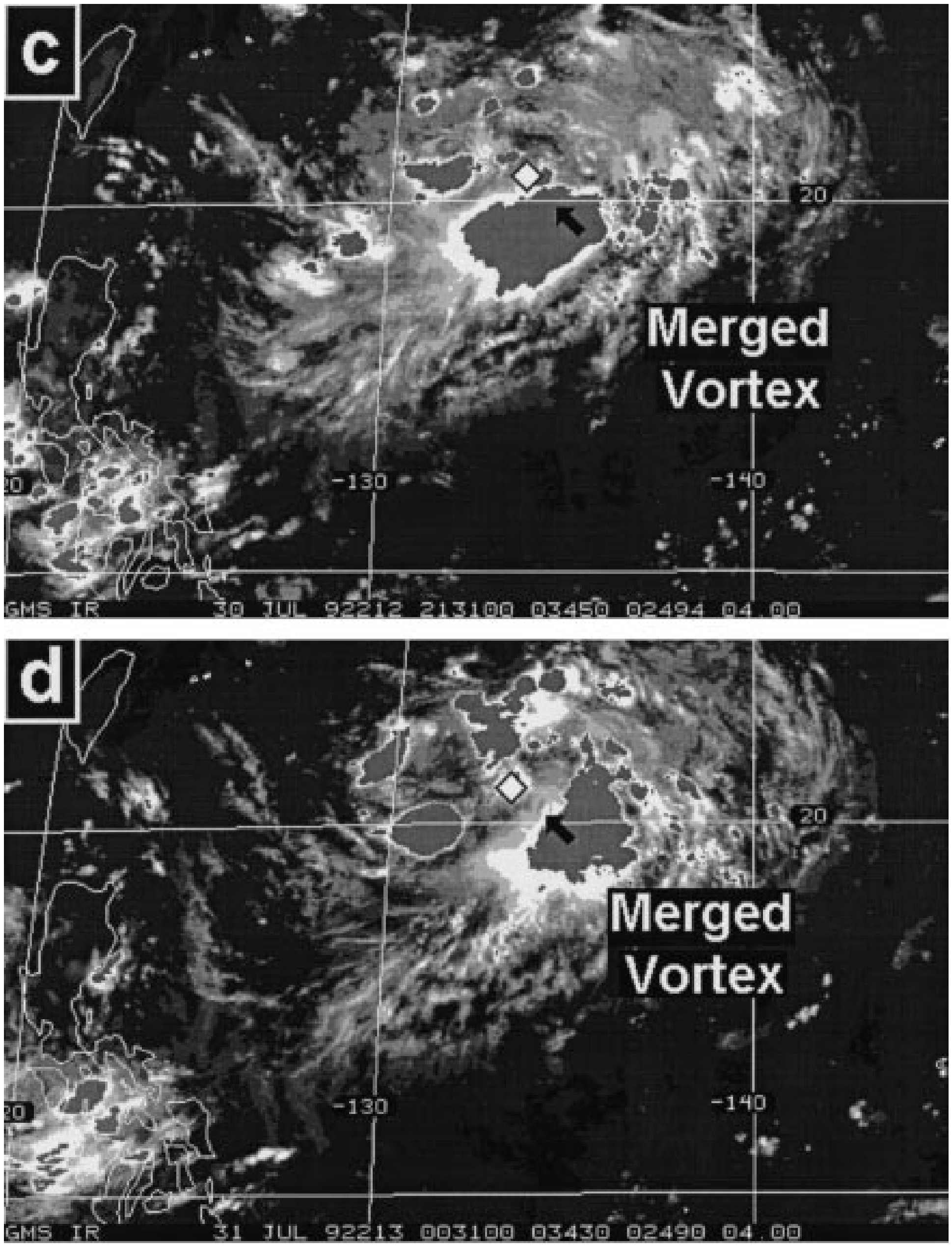

FIG. 7. (*Continued*)

confirmed the presence of several regions of enhanced midlevel vorticity (Fig. 10a) embedded in a broad, elongated, low-level cyclone (Figs. 10b,c). We suggest that the overall flow structure at 500 hPa developed as a result of mesoscale vortex development and interaction associated with the development and decay of MCSs over several days.

To support this suggestion we assumed that the mesoscale vortices labeled A and B in Fig. 10a were associated with two of the major MCSs observed in Figs. 8b and 9. An approximation of the 500-hPa vorticity field was then prepared using vortex patches following Ritchie and Holland (1993). Structural changes of the observed flow field occurring during the 12-h time of aircraft data collection were accounted for. A comparison of Figs. 10a and 11a indicates that the wind field derived from this discrete vortex patch approximation accurately reproduced the aircraft observed winds.

The analysis was used as the initial field for a contour dynamics experiment (section 2). Many modeling studies have shown that barotropic models forecast the movement and interaction of vortices to a high degree of accuracy (e.g., Fiorino and Elsberry 1989; Evans et al. 1991; Holland et al. 1991). Thus, the contour dy-

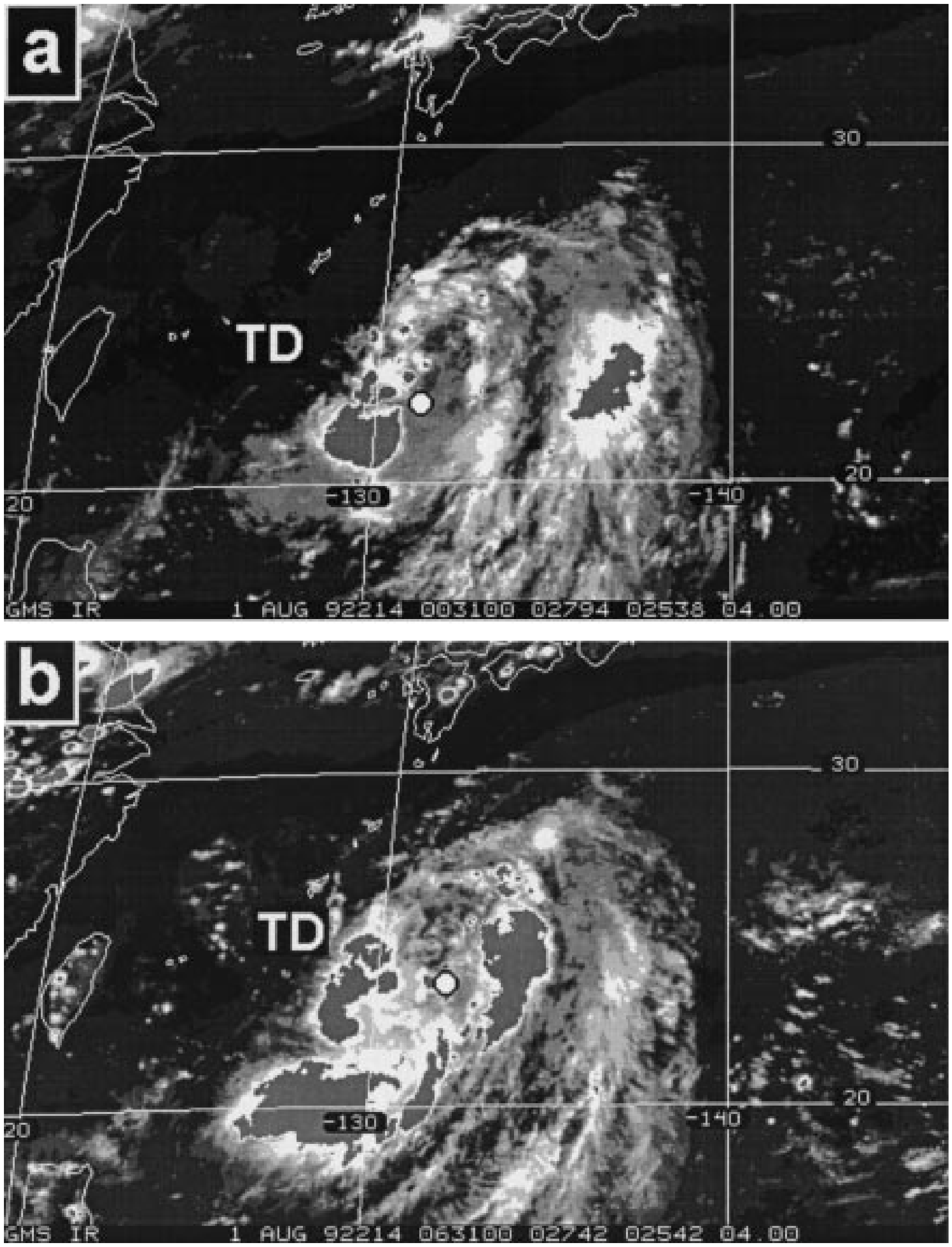

FIG. 8. Satellite imagery during the MCS–TD (tropical depression) interactions. JTWC positions are indicated by a circle: (a) 0000 UTC 1 August, (b) 0600 UTC 1 August, (c) 1200 UTC 1 August, (d) 1800 UTC 1 August, and (e) 0000 UTC 2 August.

namics model is expected to model fairly accurately the midlevel vortex motion. The model was integrated for 10 time units to produce the changes shown in Fig. 11b. Using (2), 10 normalized time units is equivalent to approximately 18 h of real time. Thus if the initial field is equivalent to the centered mission time near 0600 UTC 1 August (Fig. 8b), then the model forecast shown in Fig. 11b is for approximately 0000 UTC 2 August (Fig. 8e).

The vorticity dynamics of the model indicate that the southwestern systems should consolidate and move away from the main northern vortex, while the eastern systems should become more filamented, move toward, and merge with this vortex. This closely approximates the observed movement of the MCSs for the same time period (Figs. 8b–e), allowing for decay and reinitiation of systems. Approximate initial separation distances and interaction rotation angles are compared for the model run and the data. For the model run, if the radius of the northern patch is R, then the initial separation distance from the most northern vortex to the southwest vortex is $2.7R$. At the end of the forecast period, the separation

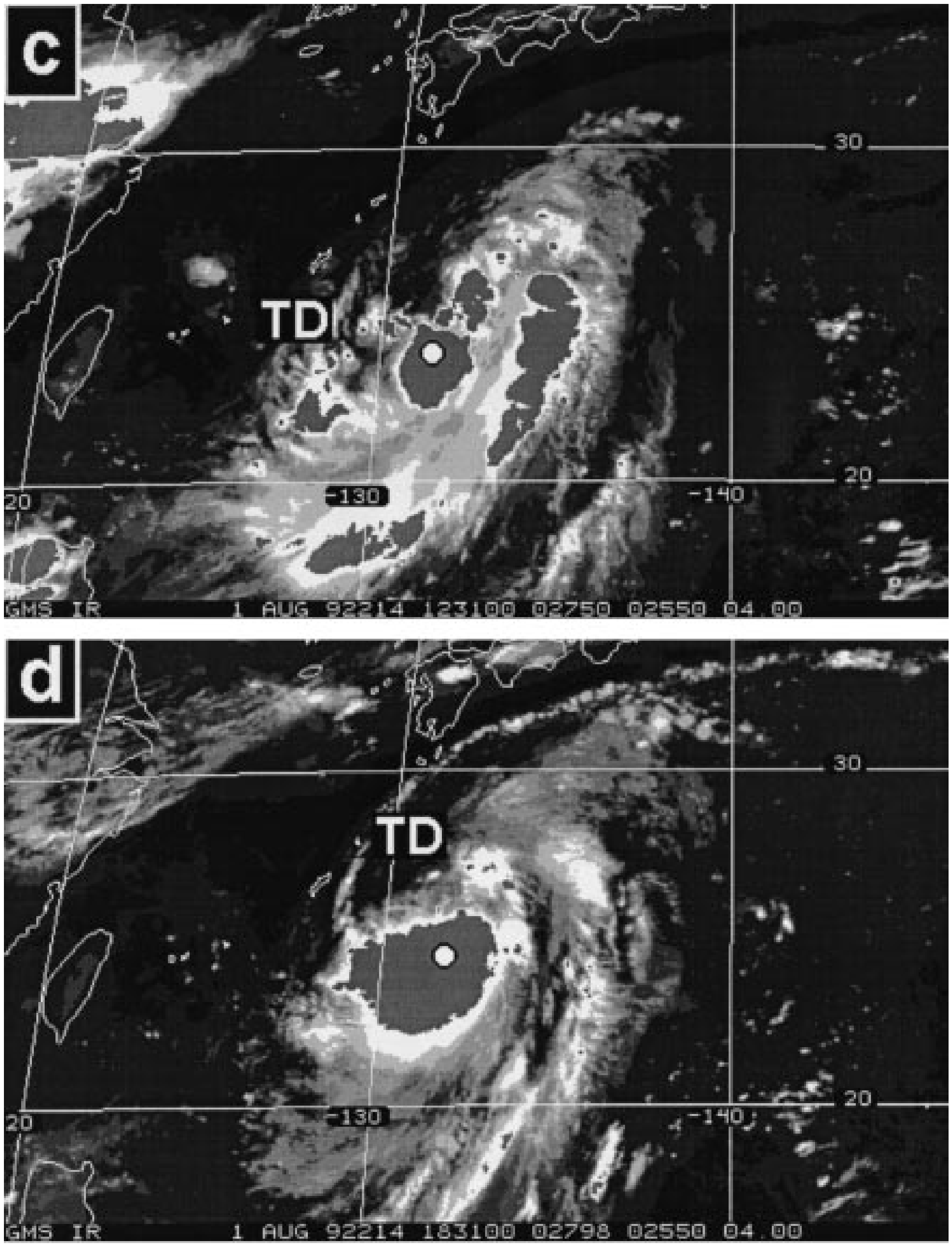

FIG. 8. (*Continued*)

distance is 3.0*R* and the systems have rotated cyclonically through an angle of 18°. From the flight-level data, initial separation distance from the most northern vortex to the southwest vortex is 3.0*R*. After 18 h, satellite imagery shows that the separation distance is 3.3*R* and the systems have rotated cyclonically through 10°.

In both the model forecast and the satellite observations, we see similar relative movement and rotation occurring. As stated earlier, as long as the model accurately forecasts vortex motion and interaction, this shows that MCS movement in satellite imagery corresponds well to the movement of their associated vortices. Recall that at this stage we are not making the assumption that all MCSs develop associated midlevel vortices. Instead we assume that once a midlevel vortex has developed, the probability that convection will redevelop in the same region as the vortex is increased. The possible mechanisms involved are discussed in detail in Fritsch et al. (1994) and some details are described in section 3. Based on this assumption, midlevel vortex motion can then be tracked in satellite imagery by using the cloud motion as a surrogate. One advantage to this is that satellite imagery is more readily available with a greater coverage than radar or aircraft data. We do

FIG. 8. (*Continued*)

note that the vortices and cloud clusters may not have tracked together as well as we might hope. Without continual forcing, MCS gradually decay, whereas their associated vortices live on. Vortices from previous convective systems may have been responsible for developing new convective systems. In particular, evidence of redevelopment of convective clusters near the advective path of some vortices is shown in Fig. 9. Thus, using satellites to track midlevel vortices becomes a matter of predicting the path of the vortex then looking for redevelopment of the MCS as confirmation of this path.

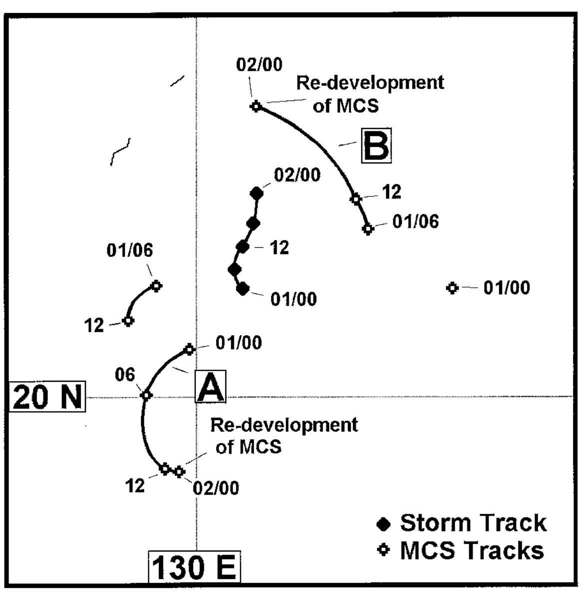

FIG. 9. Tracks of the storm and three MCSs during the interactions illustrated in the satellite imagery in Fig. 8. The A and B label corresponding vortices in Fig. 10. All times are UTC.

d. Potential impact of mesoscale vortices on the storm development

We have established that the development of Irving was characterized by the development of a number of MCSs and that those for which aircraft reconnaissance data were available had associated strong mesoscale vortices in the midtroposphere. The barotropic results in the previous section demonstrate that vortex interaction may account for most of the MCS movement observed in the satellite imagery if, as we presume, each MCS acts as a flag for an associated vortex. However, the contour dynamics model does not address questions regarding the vertical changes occurring during the interactions. Thus, the question naturally arises as to the role, if any, that these systems have in the cyclone development.

The aircraft observations show that these midlevel mesoscale features contain considerable vertical structure with height and cannot be considered approximately barotropic, as was the case in the last section. They have maximum amplitude near 500 hPa, whereas the larger-scale cyclonic circulation within which they developed has maximum amplitude near the surface. The effects of vertical influences on ''dynamically small'' vortices, that is, vortices with a horizontal scale, $L \ll L_R$, can be examined through the modified Rossby–Burger–Prandtl relationship (Hoskins et al. 1985)

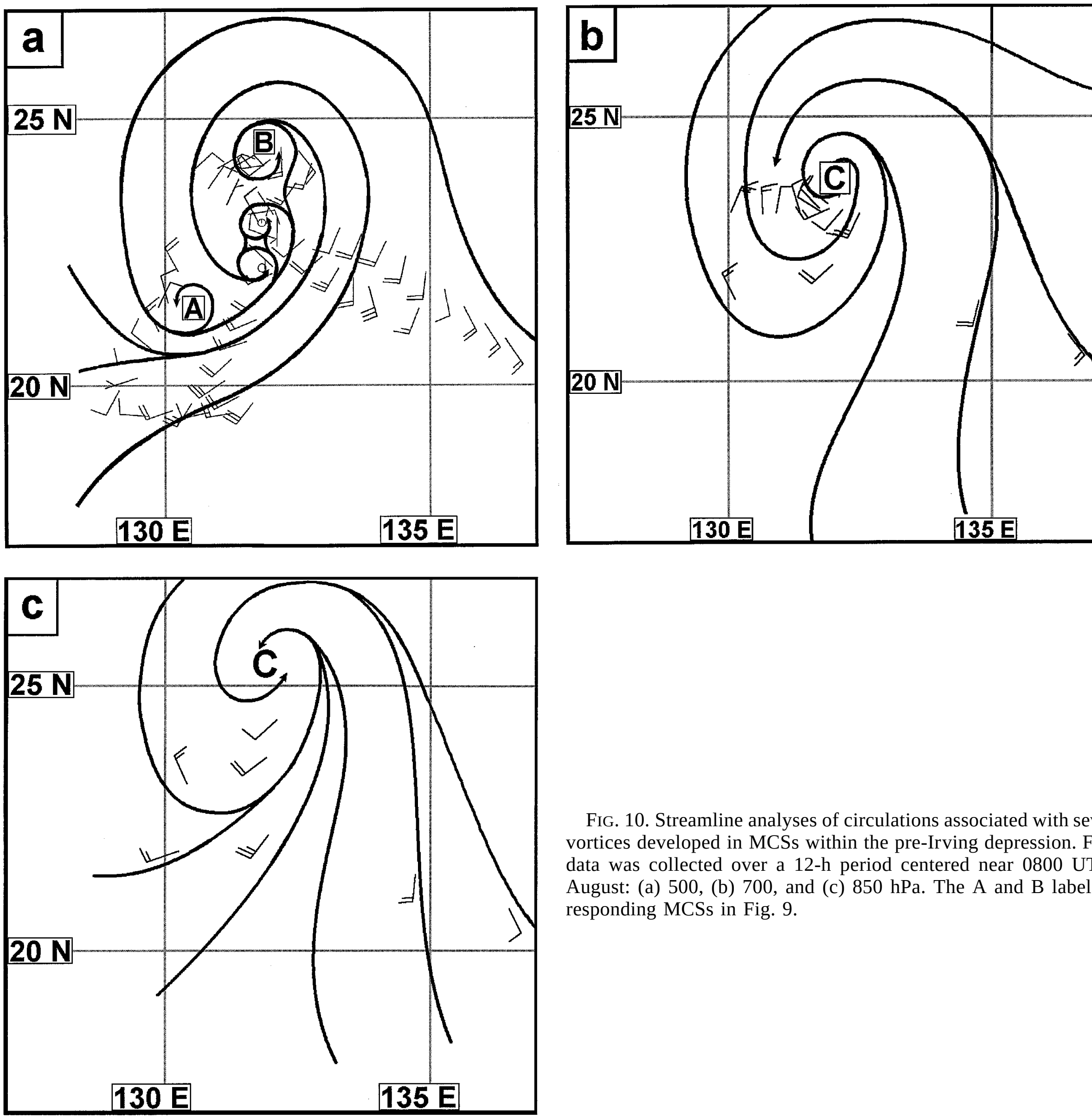

FIG. 10. Streamline analyses of circulations associated with several vortices developed in MCSs within the pre-Irving depression. Flight data was collected over a 12-h period centered near 0800 UTC 1 August: (a) 500, (b) 700, and (c) 850 hPa. The A and B label corresponding MCSs in Fig. 9.

$$D = \frac{(f_{\mathrm{loc}}\zeta_a)^{1/2}L}{N}. \quad (5)$$

Here D is the vertical influence of a PV perturbation, L is the horizontal scale of the perturbation, f_{loc} is a local background rotation term, ζ_a is absolute vorticity (taking Coriolis into account), and N is the Brunt–Väisälä frequency, a measure of the thermal stratification of the atmosphere. Thus, a vortex, initially restricted to the midlevels, can develop vertically by increasing either the horizontal size of its PV perturbation or local background rotation or by decreasing its local vertical temperature gradient, which partitions the PV into its relative vorticity component (4). Since the interaction and merger of two-dimensional vortices results in a vortex of greater size (e.g., Zabusky et al. 1979; Ritchie and Holland 1993), a probable consequence of merging midlevel vortices is a downward growth of the vorticity as a direct consequence of (5). The addition of significant background rotation in the form of large-scale, low-level circulations can also increase the downward development of the midlevel vortices according to (5). This includes nonlinear feedbacks between the midlevel and large-scale circulations, which are difficult to predict but are assumed to have important consequences for growth of large-scale circulations to dynamically smaller (tropical cyclone) scales.

The scale interactions observed between these mid- and lower-level vortices during the formation of Typhoon Irving are investigated using the primitive equa-

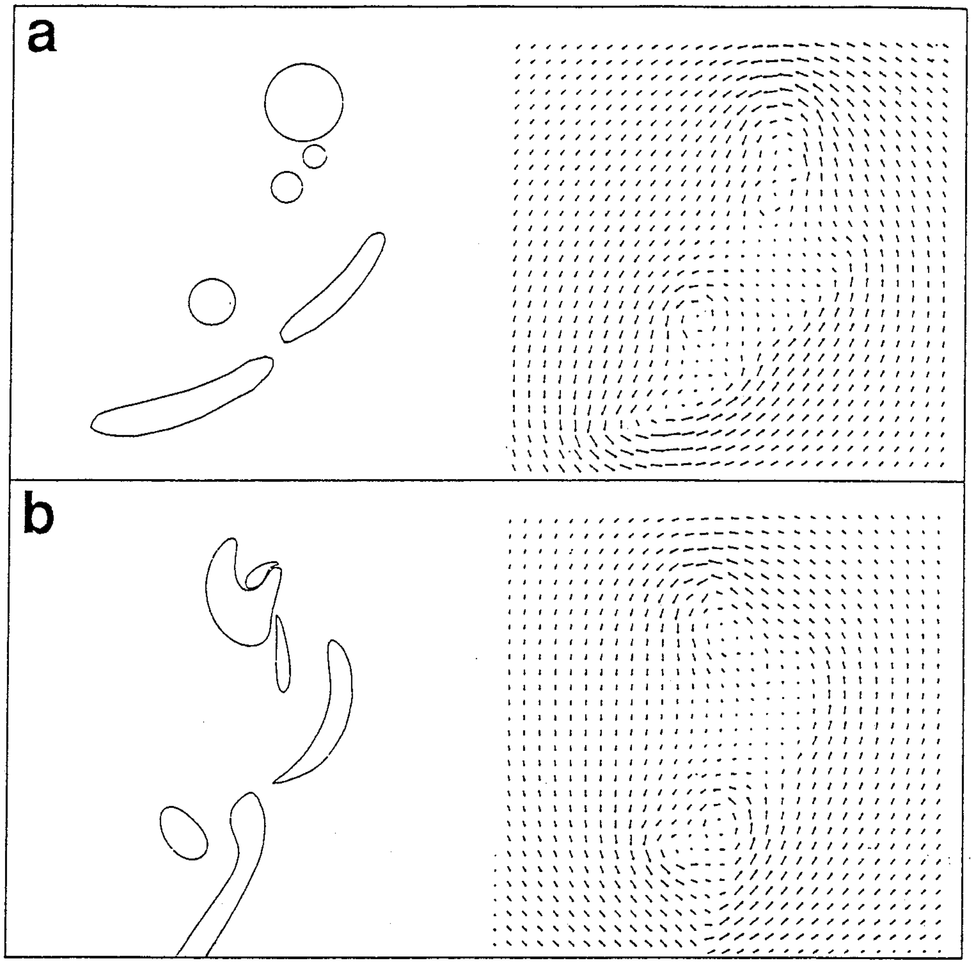

FIG. 11. Vortex patches and associated wind field for 2D simulation of MCS interactions in Typhoon Irving: (a) $t = 0$ and (b) $t = 10$. All space and time variables are nondimensional.

tions model described in section 2 (Wang and Holland 1995) on an f plane, valid at 15°N. It has been reported (e.g., Raymond and Jiang 1990) that development of convection through a midlevel potential vorticity maximum will lead to a rapid flux of PV to the surface. However, there is no evidence that this occurs in Irving, where the vortices seem to be associated with midlevel stratiform cloud. Hence, we assume that the primary dynamical response to organized mesoscale convection in this situation is the development of midlevel mesoscale vortices. In order that we may investigate a hypothesis that simple vortex dynamics could be a significant component of the development of Irving, all diabatic heating, moist processes, and frictional effects are turned off in the model.

Simple interactions between two midlevel, vertically constrained vortices are described in detail in Ritchie (1995). Here we will briefly summarize results relevant to the present discussion. Two vortices located within a critical separation distance (dependent on the size, strength, tangential wind profile, and background flow) will merge, forming a new vortex of greater horizontal and vertical size (Figs. 12a,b). Interestingly, the extension in the vertical of the new vortex occurs through the advection of mass out of the layer of interaction and into surrounding layers to conserve energy in the domain. Some mass is also lost to horizontal filamentation and eventual dissipation.

When the invariant background rotation is increased in the system, merger is more efficient, allowing less mass loss to horizontal filamentation. Thus energy conservation requires more mass to be evacuated out of the layer of interaction and into surrounding layers. This results in an even larger increase of the vortex through vertical layers. The vortex has reached the surface by 120 h of integration (Fig. 13) and, in fact, had reached it by 72 h. In this experiment the invariant background has a vertical structure similar to that of the large-scale surface circulation associated with Irving's development, that is, maximum at the surface and decreasing with height to about 500 hPa.

The interactions described in section 4c involved one

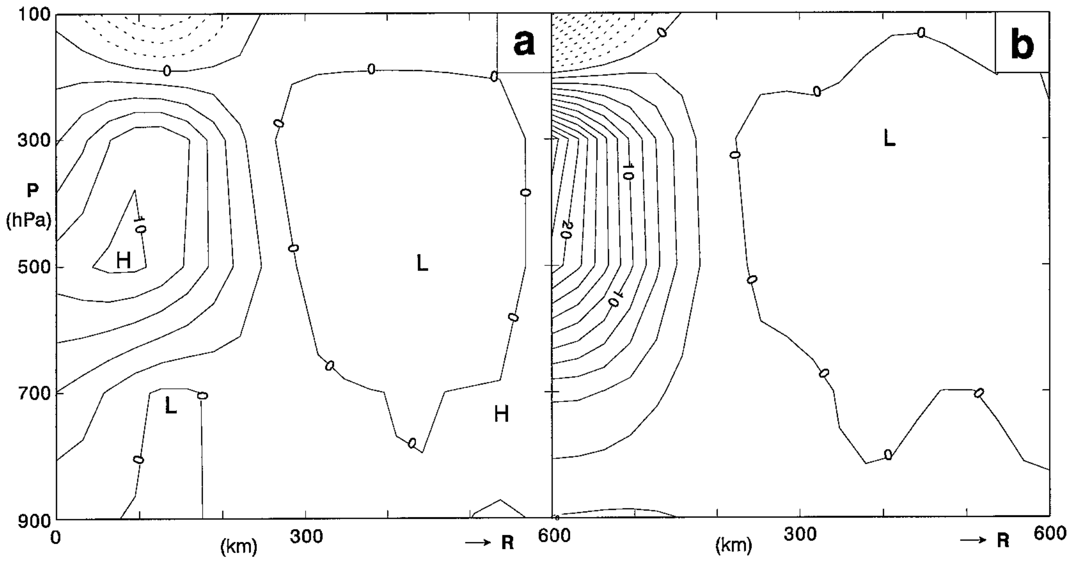

FIG. 12. Baroclinic model simulation of the interaction between two equal midlevel vortices: (a) Azimuthal average of relative vorticity at the initial time. (b) Azimuthal average of relative vorticity for the merged vortex at 72 h of integration. Contour intervals are 2.0×10^{-5} s^{-1}.

or more small mesoscale vortices with maximum amplitude in the midlevels near the base of stratiform cloud decks and a noninvariant, synoptic-scale vortex with maximum amplitude near the surface. This is simulated by constructing an idealized situation consisting of the relative vorticity field shown in Fig. 14a where two midlevel, PV perturbations are placed close to a larger-scale, low-level vortex. The mass and wind fields are derived as discussed in section 2. Here the idea is to investigate the impact of midlevel vortices on the low-level, large-scale circulation. As stated earlier, previous studies (Ritchie 1995) maintained an invariant large-scale rotation. The results are shown in Figs. 14b,c.

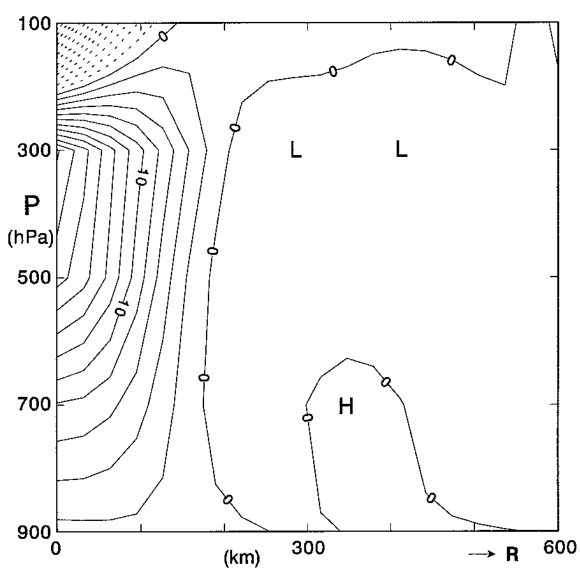

FIG. 13. Azimuthal average of relative vorticity at 120 h of integration for the interaction in Fig. 12, with an additional invariant background rotation similar to a monsoon trough structure. Contour intervals are 2.0×10^{-5} s^{-1}.

After 120 h of integration, the vortices have combined, aligning in the vertical to form a single, integrated vortex that extends through the troposphere farther than any of the original vortices alone (Fig. 14b). The new vortex has strengthened at the surface, increasing its vorticity by more than 20% over the low-level vortex alone in a similar period (not shown). Figure 14c shows the difference in the 120-h vorticity field from the initial configuration, demonstrating both the increase in mid- to upper-tropospheric vorticity above the low-level vortex, as well as the increase in vorticity at the surface.

The vortices initially interact only in a small layer of the total vertical domain. The two midlevel vortices merge, forming a new midlevel vortex that is larger and deeper and centered over the center of the low-level vortex. The larger, low-level vortex provides a region of enhanced lower-level rotation of about 23×10^{-5} s^{-1} or $6f$. Ritchie (1995) showed that such an environment induces strong downward growth of midlevel vortices. This is also predicted by the $(f_{\text{loc}}\,\zeta_a)^{1/2}$ term in (5). Thus, the midlevel vorticity extends toward the surface due to the enhanced lower-level, large-scale rotation and begins to interact with the low-level vorticity that is initially unaffected by the merger. In effect, the midlevel vortex extension becomes a core about which the more loosely organized low-level vortex can wrap around, becoming tighter and stronger. Thus, the main difference between this experiment and the one reported in Ritchie (1995) with an invariant background rotation is the feedback of the midlevel vortex onto the larger, surface vortex. The wrapping up of the surface vortex initiates

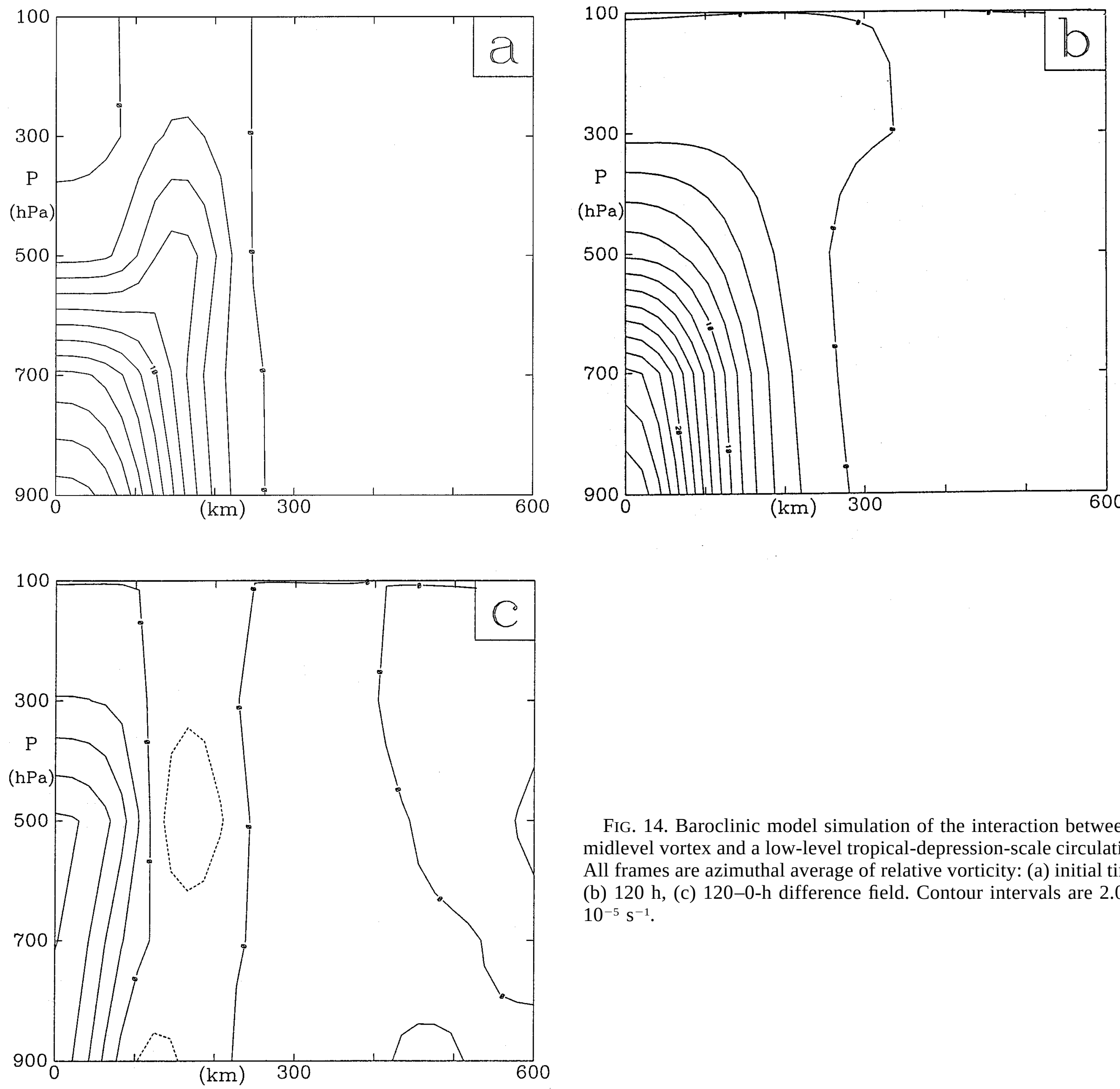

FIG. 14. Baroclinic model simulation of the interaction between a midlevel vortex and a low-level tropical-depression-scale circulation. All frames are azimuthal average of relative vorticity: (a) initial time, (b) 120 h, (c) 120–0-h difference field. Contour intervals are 2.0 $\times 10^{-5}$ s^{-1}.

convergence toward the vortex center. In a system with forcing included, this convergence would enhance upward motion and reinitiation of convection, development of MCSs, and midlevel vortices. Once begun, the system can, in theory, maintain itself even in a hostile environment.

5. Summary and conclusions

We have investigated the synoptic and mesoscale conditions during the preformation and formation periods of Typhoon Irving using synoptic and research aircraft data. It was found that although generally hostile environmental conditions existed prior to Irving's development, interactions between synoptic-scale cyclonic circulations enhanced the local background vorticity and created a locally more favorable area for cyclogenesis. A subsequent interaction between the low-level merged circulation and a midlevel cyclonic feature associated with a TUTT resulted in formation of a tropical-depression-strength vortex. It may be argued that the observed depression had developed directly from the TUTT-induced, midlevel circulation. The convection associated with this circulation is the feature that JTWC tracked prior to tropical depression (TD) classification. However, the large-scale analyses showed no evidence of a low-level circulation near the midlevel vortex. The only low-level feature in the region was the merged cyclone, located farther to the south and west from the TUTT-induced midlevel cyclone.

During this synoptic-scale interaction period, midlevel mesoscale vortices were observed within the circulation boundaries of the low-level circulations. Evidence exists that some of these vortices developed within weakly organized MCSs (McKinley 1992). Their dynamic role, if any, during this period is not well understood, but baroclinic model results suggest that their interaction with the low-level circulation may be serving to strengthen the circulation at the midlevels.

After the low-level circulation reached TD strength, more vigorous MCS activity was observed with associated midlevel vortices. These were observed to interact with the main circulation, rotating and sometimes merging. Contour dynamics experiments show that simulated vortices on the east side of the system merged with the main vortex, while vortices on the west and south sides moved away. Comparison with satellite imagery, using MCSs as tracers for midlevel vortices, show a similar pattern. The simple baroclinic model results suggest that the merging vortices contributed midlevel vorticity to the TD and helped to tighten up the low-level circulation. Thus, we see that the initial development of Irving required a cooperation between the large-scale and the mesoscale dynamics. The development was modulated by vortex interactions in the gradient to 700-hPa layer and at 500 hPa, where frequent MCS activity resulted in continual mesoscale-vortex formation.

We especially note that we have neglected the moist dynamics other than to state that we believe the primary response to organized mesoscale convection is to concentrate PV in the midtropospheric stratiform regions. The direct effects of diabatic heating on these MCS-generated vortices is not well understood. Previous studies (e.g., Fritsch et al. 1994) have argued that the development of convection near the center of such vortices serves to enhance the midlevel vortex. Other studies (e.g., Raymond and Jiang 1990; Hertenstein and Schubert 1991) argue that convection acts as a mechanism to flux PV near the surface, destroying the midlevel structure. All agree that stratiform heating serves to maintain and enhance midlevel PV structure. Questions arise as to the mechanism that retriggers convection near a midlevel PV anomaly over the tropical oceans. All these are questions that merit further study.

We cannot make generalizations for all cyclogenesis cases from this one case study. Nevertheless, we hypothesize that vortex interaction modulates most tropical cyclogenesis cases, at least in the western Pacific region where cyclogenesis processes are dominated by vigorous MCS activity. Other work to be reported will focus on the dynamics and growth of interacting midlevel vortices and their role in tropical cyclogenesis as a general mechanism for formation.

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