Large Intensity Changes in Tropical Cyclones: A Case Study of Supertyphoon Flo during TCM-90

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ABSTRACT

A unique dataset, recorded during the rapid intensification and rapid decay of Typhoon Flo, is analyzed to isolate associated environmental conditions and key physical processes. This case occurred during the Tropical Cyclone Motion (TCM-90) field experiment with enhanced observations, especially in the upper troposphere beyond about 300 km. These data have been analyzed with a four-dimensional data assimilation technique and a multiquadric interpolation technique. While both the ocean thermal structure and vertical wind shear are favorable, they do not explain the rapid intensification or the rapid decay. A preconditioning phase is defined in which several interrelated factors combine to create favorable conditions: (i) a cyclonic wind burst occurs at 200 mb, (ii) vertical wind shear between 300 and 150 mb decreases 35%, (iii) the warm core is displaced upward, and (iv) 200-mb outflow becomes larger in the 400–1200-km radial band, while a layer of inflow develops below this outflow. These conditions appear to be forced by eddy flux convergence (EFC) of angular momentum, which appears to act in a catalyst function as proposed by Pfeffer and colleagues, because the EFC then decreases to small values during the rapid intensification stage. Similarly, the outer secondary circulation decreases during this stage, so that the rapid intensification appears to be an internal (within 300 km) adjustment process that is perhaps triggered during the preconditioning phase.

Rapid decay occurred over open ocean when the environmental factors of ocean thermal structure, and vertical wind shear, positive 200-mb EFC, and vigorous outflow into the midlatitudes appear favorable. However, the EFC extending down to 500 mb and inducing a second shallower secondary circulation is hypothesized to account for the rapid decay by leading to a less efficient energy conversion.

1. Introduction

The problems of understanding and predicting tropical cyclone (TC) intensity have been highlighted in the American Meteorological Society’s Symposium on Tropical Cyclone Intensity Change in Phoenix, Arizona, during January 1998. Thirteen keynote papers and 31 other contributions summarized the present state of knowledge on TC intensity change. Other than being a fascinating and challenging scientific problem, an accurate intensity prediction is an important factor in the warnings of landfalling TCs (Marks and Shay 1998; Elsberry and Marks 1998). In general, slow intensification of TCs is predicted by the U.S. National Hurricane Center reasonably well (M. DeMaria 1997, personal communication), with mean 24-h intensity forecast errors of about 10 kt (5 m s$^{-1}$). However, rapid intensity changes are not predicted well, with 5% of the 24-h intensity forecasts with errors exceeding 40 kt (20 m s$^{-1}$). Such large errors in a TC intensifying close to a coast could mean that the public would not have time to prepare.

It is particularly important to forecast accurately the most intense TCs, since it has been estimated (Pielke and Landsea 1998) that 80% of the damage in landfalling U.S. hurricanes is caused by TCs with Saffir–Simpson categories of 3–5. It has been known for some time that these most intense TCs will have experienced at least one period of rapid intensification. For example, Holland (1984) found intense TCs in the Australian region typically intensify at twice the rate, and for three times as long, as minimal hurricane-strength systems. Thus, the focus should be on understanding and predicting these periods of rapid intensification.

It is also of interest to understand and predict the rapid decreases in TC intensity. The case of Hurricane Opal that rapidly weakened just prior to striking the Gulf of Mexico coast is the focus of many of the papers in the TC intensity change symposium. Although it is fortunate in terms of the potential damage that is averted, credibility of the forecasts can suffer if such large intensity decreases are not predicted well.

The favorable environmental factors involved in TC intensification are relatively well established (Frank
1987; McBride 1995). More recent summaries are given in the TC intensity change symposium, and the reader is referred to that preprint volume for details. In general terms, higher sea surface temperatures (SSTs) are considered favorable because the energy source for the TC is the fluxes of heat and moisture from the warm tropical ocean [see keynote papers by Emanuel (1998) and Black and Shay (1998)].

Large environmental vertical wind shear is considered unfavorable for intensification, and is believed to be a major factor in the rapid decay of TCs. For example, values of vertical shear above certain threshold values are believed to prevent TC intensification beyond the tropical storm (35–63 kt, 17–32 m s\(^{-1}\)) stage [see keynote papers by DeMaria and Huber (1998) and Jones (1998)]. However, Molinari (1998) states that storms have intensified despite high vertical shear during interactions with upper-tropospheric troughs. Elsberry and Jeffries (1996) caution that vertical shear estimates from operational wind analyses in the Tropics may be faulty because of data-sparse problems over developing TCs, and Elsberry (1998) suggests these questionable vertical shear estimates may be a primary factor in poor TC intensity predictions with statistical equations. Less agreement exists as to how the TC intensity is related to vertical shears below the threshold value. Zehr (1998) takes the rather extreme position that vertical shear is the only atmospheric environmental influence that determines the TC intensity! DeMaria and Huber (1998) propose that the physical processes involved in TC intensity may change with the magnitude of the environmental vertical shear and the existing storm structure.

A particularly complex and controversial topic is the influence of adjacent midlatitude or tropical troughs in the upper troposphere on TC intensity changes. Studies back to Pfeffer (1958) have suggested that a proper juxtaposition with an adjacent trough will lead to an eddy angular momentum flux convergence into the TC in the upper troposphere. Molinari (1998) considers this TC–trough interaction to be an example of a “good trough” for TC intensification. Molinari et al. (1995, 1998) propose superposition of a positive potential vorticity feature in the upper troposphere with a similar horizontal scale as the TC potential vorticity may be an essential feature in some cases of rapid deepening. As Wu and Cheng (1998) demonstrate using a potential vorticity (PV) inversion technique, and as will be illustrated from the analyses below, evidence for a PV superposition is lacking in the case of Typhoon Flo. Juxtaposition of the trough may also increase the vertical shear over the TC, and this would be a “bad trough” effect on TC intensity. In a series of modeling studies culminating in a recent paper (Challa et al. 1998), Pfeffer and colleagues have studied environmental forcing of TCs. This group (and several others) have simulated TC formation/intensification in response to strong and persistent environmental forcing. DeMaria et al. (1993) found the large eddy angular momentum flux events were sometimes, but not always, correlated with future TC intensity changes. Some factors may be the TC stage at the time of environmental forcing event; how long, and at what radius and over what depth, the eddy momentum flux occurs; and the intermediate storm structure that determines the inner-core response to a remote (large radius) forcing event. Given these many factors by which environmental forcing may be involved in TC intensification, perhaps it is not surprising that diverse opinions exist as to the role of environmental forcing.

A fourth factor in TC intensification is the internal dynamics of how the heat/moisture gained from the ocean, and transferred to the atmospheric boundary layer, is released. Clearly, factors such as the radial distribution of clouds with various depths are important. One fairly well accepted internal dynamic mechanism for TC intensity changes is the contracting eyewall cycles of Willoughby et al. (1982). As the outer eyewall contracts inward, the maximum surface wind (intensity) associated with the inner eyewall decreases. As the outer eyewall replaces the inner eyewall, the intensity again increases (see Willoughby 1995 for examples and a complete description). Although these contracting cycles seem to occur relatively frequently in TCs with an intensity greater than 50 m s\(^{-1}\), the mechanism(s) that triggers the cycles is not known. As Holland and Wang (1998) note, these internal dynamic processes have specific timescales and life cycles, and these dictate to some extent the adjustment of the TC to external forcing, and thus affect the rate of intensification.

In summary, four nonlinearly interactive processes (ocean thermal structure changes, atmospheric boundary layer and convective transports, and external factors such as upper-tropospheric static stability and eddy angular momentum transports) are involved in TC intensity changes. Although many numerical simulations have been made to explore the sensitivity to these various factors, relatively few observational studies of rapid intensity changes have been published. This is particularly true in the western North Pacific region.

The objective of this study is to describe the rapid intensity changes during the life cycle of Supertyphoon Flo (intensity > 130 kt, 65 m s\(^{-1}\)) that occurred during the Tropical Cyclone Motion (TCM-90) field experiment (Elsberry 1990; Elsberry et al. 1990). Although the purpose of this experiment was to collect a comprehensive dataset to study western North Pacific TC motion, the dataset will be used here to study TC intensity changes. In particular, the rapid intensity changes that led to Flo having achieved supertyphoon intensity (Fig. 1a) will be contrasted with the simultaneous intensification of Typhoon Ed, which did not exceed 90 kt or 46 m s\(^{-1}\) (Fig. 1b). Maximum intensity of Flo was achieved after a long track toward the northwest. Later in the life cycle of Flo, a rapid decay occurred over open ocean prior to landfall on Japan. Typhoon Ed had a long westward track along 20°N with only a slow
intensification rate. After brushing the northern tip of Luzon Island, Philippines, maximum intensity was achieved over the South China Sea.

In contrast to the Challa et al. (1998) comparisons of simulations of a supertyphoon development and a non-developing TC, this observational study searches for factors that distinguish rapid intensity changes in Flo from those in a nominal typhoon development. Whereas Merrill and Velden (1996) have previously carefully analyzed the evolution of the outflow channels during the rapid intensification and decay of Supertyphoon Flo, this study will describe some of the physical processes associated with these intensity changes. The two independent analyses of the TCM-90 dataset that are used in this study will be described in section 2.

2. Methodology

a. TCM-90 dataset

The TCM-90 field experiment was conducted in the western North Pacific during August and September 1990. As discussed in Elsberry (1990), this experiment was in conjunction with three other TC initiatives sponsored by the Economic and Social Commission for Asia and the Pacific/World Meteorological Organization, the former Soviet Union, and Taiwan. The combined experiments produced an enhanced dataset: (i) 6-h rawinsonde coverage during intensive observing periods (IOPs); (ii) additional over-ocean rawinsondes from four Soviet Union ships stationed along 20°N; (iii) special observing platforms such as radar wind profilers, drifting buoys, and especially three flights of the National Aeronautics and Space Administration (NASA) DC-8 aircraft into Typhoon Flo; and (iv) postexperiment processing of the Geostationary Meteorological Satellite cloud-drift winds by the Cooperative Institute for Meteorological Satellite Studies at the University of Wisconsin, with careful manual assignments of altitudes based on the temperatures from the enhanced rawinsonde network. A complete list of data collected during each IOP is given in Harr et al. (1991). Examples of the observations during 0600 UTC periods, except with special DC-8 observations, are given later (Figs. 13 and 14). Many of the key conclusions in the following analysis are based on the upper-tropospheric observations, especially the angular momentum flux calculations and the outflow characteristics.

b. Two objective analyses

The National Centers for Environmental Prediction (NCEP) Environmental Modeling Center prepared a four-dimensional data assimilation (4DDA) analysis of the TCM-90 dataset (Rogers et al. 1992). Because the 4DDA technique blends the observations with the 6-h forecast of the NCEP mesoscale (50-km horizontal resolution) Eta Model, efficient incorporation of asynoptic data, dynamical consistency within and between levels, and time consistency are expected advantages. However, the technique depends on prescribed statistical relationships between the observations and the Eta Model forecast values. It is certainly a question whether these statistical relationships are valid in regions with small Coriolis values and low inertial stability such as the outflow layer of a typhoon. To avoid bias in the diagnostic studies with these 4DDA analyses, no synthetic TC observations were included to define the position and structure. In the absence of observations near the center, the Eta Model 6-h forecast is likely to be the primary de-
terminant of the TC position and structure. Another potential problem is the model parameterization of latent heat release in the TC, which might constrain the observations to a thermal structure that differs from the real atmosphere.

As an alternative, a multiquadric (MQ) interpolation technique is used to convert irregularly spaced observations to a regular grid. This TCM-90 application follows from the Nuss and Titley (1994) MQ implementation, which is a linear combination of circular hyperboloids. These functions are fit closely to the observations while accounting for the likely errors in the different observation types. These functions have continuous derivatives and they may be interpolated to the regular grid with an empirically derived smoothing parameter, which determines how sharply the gradient of the circular hyperboloid changes near each observation. Details of the parameter specifications, observational error specifications, and tests of the scheme are given in Titley (1998).

Several special aspects of the MQ application to the TCM-90 dataset are briefly summarized here. First, the analysis was of increments that are the differences between the observations and values interpolated with a B spline to the observation location in the Navy Operational Global Atmospheric Prediction System (NOGAPS) analyses done in real time. Although these 2.5° latitude–longitude NOGAPS analyses are relatively coarse, they provide a good dynamical synoptic basis for a higher-resolution analysis in the region of TCM-90 observations. Zero-increment bogus values were inserted at grid points at least 1500 km from a real observation, which constrains the MQ analysis to return to the NOGAPS analysis. Since each MQ analysis is on a two-dimensional pressure surface, the NOGAPS analyses provide a background vertical structure that is in hydrostatic balance. It is emphasized that each MQ analysis is a “cold start,” that is, the NOGAPS analysis is a background for calculating the increments, but the MQ analysis is not a blending of observations and a short-term numerical model forecast as in the 4DDA analyses. Whereas the MQ analysis will not be as model physics dependent as the 4DDA analysis, it will not have the benefit of time continuity and spreading the information into data-sparse areas that comes from the analysis–forecast–analysis cycle of the 4DDA, except as is provided by the NOGAPS assimilation system.

The MQ analyses of geopotential heights are geostrophically constrained north of 35°N, unconstrained south of 25°N, and a linear blend between 25° and 35°N. The MQ analyses of temperature are not constrained by winds or heights. Gradient winds based on the MQ analyzed height field are the first guess for the wind analyses north of 35°N and within 300 km of tropical cyclones. Winds and heights within 300 km of a TC are replaced by an azimuthal average in radial increments of 50 km in a moving cylindrical coordinate system, and blended with the Cartesian coordinate system between 300 and 600 km of a TC. This azimuthal averaging of the height field occurs during the first guess construction, and the averaging of the wind field occurs after the final MQ analysis. The reason for the azimuthal averaging near TCs is to avoid oval-shaped fields where the MQ analysis fit closely to an observation near a TC is not matched with well-distributed observations around the TC to maintain the typical near symmetry of fields. Details of these procedures and examples of the final MQ analyses near TCs are given in Titley (1998).

These special procedures for the MQ analyses are designed to produce wind and height analyses fit to the observations similar to those analyzed by a skilled human analyst. These MQ analyses will generally be selected for presentation in the following sections because the MQ technique fits the observations more closely. An advantage of the 4DDA analyses is that they are available for other TCM-90 storms besides Flo and Ed.

3. Preconditioning phase of Flo

The various favorable environmental factors for TC intensification described in the introduction are first examined for Flo. First, no ocean-related evidence could be found distinguishing the more rapid intensification of Flo relative to the conditions existing in Ed. Both typhoons intensified over a broad region with SSTs of about 28.0°C–28.5°C (not shown), which are within 0.5°C of climatological values. The Philippine Sea region in which both TCs formed typically has ocean mixed layer depths exceeding 100 m, so that the heat content is sufficient for intensification even if either Flo or Ed had been moving slowly. No correlation of intensity change with the passage of either Flo or Ed over a warm-core ocean eddy (as in the frequently studied Hurricane Opal case) was evident.

Calculations of vertical wind shear are possible with the TCM-90 analyses each 50 mb. Deep tropospheric, low-level, middlelevel, and upper-level differences in wind vectors have been calculated (Titley 1998) and contrasted between Flo and Ed. Averages over a variety of horizontal domains relative to the center have been calculated. Considering the minimum data coverage near the center, this region has been excluded. Although an average vertical shear over the 200–600-km radial band is presented here, other radial bands have similar variations in time (Titley 1998).

Eddy fluxes of angular momentum as in Molinari et al. (1995) have been calculated at each 50 km in radius and 50 mb in depth for a moving coordinate system using the observed translation speeds (Figs. 1a and 1b). Eddy flux convergences are then calculated at various radii and converted to equivalent tangential wind speed changes (m s⁻¹ day⁻¹) averaged over the cylindrical volume within that radius.
a. Preconditioning conceptual model

A conceptual model summarizing the differences during the preconditioning phase of Flo relative to that of Ed is shown in Fig. 2. Each of the special aspects of Flo will be described with selected diagrams to illustrate this conceptual model. More detailed analyses are available in Titley (1998).

The left two panels of Fig. 2 illustrate the thermal structure differences and emphasize the upward displacement of a warm anomaly in Flo that did not occur in Ed. These thermal structure differences may be illustrated by azimuthal-averaged temperatures (Fig. 3) in a storm-centered coordinate system for six times at which the storms were of similar intensities. For five of these times between 0000 UTC 14 September and 1800 UTC 15 September, the storms had identical intensities, and the intensity of Ed was only slightly (5 kt, 2.5 m s^{-1}) greater at the sixth time. Whereas both Flo and Ed intensified from 45 kt (22.5 m s^{-1}) to 85 kt (42.5 m s^{-1}), Flo (Ed) achieved this intensity change over only 36 h (96 h). It is emphasized that all of the times related to Flo were prior to the rapid intensification, so that this is considered to be a "preconditioning effect." Although these inner-region temperature differences are calculated from the 4DDA analyses that may have some Eta Model dependencies, Ed was moving westward along 20°N where four Soviet Union ships were providing rawinsondes each 6 h and Flo passed near the Saipan and Guam rawinsondes. Since the key features in Fig. 3 are consistent in time, these temperature differences are believed to be real and not an artifact of the Eta Model.

These temperature differences indicate the inner core of Flo is less statically stable than that of Ed, which implies the midtropospheric thermal structure of Flo is more conducive to sustained convection near the center. However, the upper troposphere above Flo between 150 and 100 mb is 1°–2°C higher than above Ed. That is, these composited six analyses are consistent with Gray’s (1992) hypothesis of strong, intensifying storms having greater convective instability in the midtroposphere and higher temperatures aloft. These temperature differences are summarized in the left panels of Fig. 2.
The middle panels of Fig. 2 summarize the corresponding tangential wind structure changes. Traditionally, vertical wind shear has been calculated between 850 and 200 mb because these two levels typically have the best wind coverage from low- and upper-level cloud-drift winds. The 6-h estimates of 850–200-mb shear (Fig. 4a) are somewhat noisy. If a 1–2–1 filter is applied, the MQ estimates at 0000 and 0600 UTC 15 September are about 6 m s$^{-1}$, which is nearly the same value as at 0600 and 1200 UTC 13 September, although smaller shear values are estimated between these times. Whereas the 4DDA shear estimates at the end of the preconditioning period are also about 6 m s$^{-1}$, the initial values are higher and would suggest an initial decrease in shear. Accepting the MQ estimates, the middle schematics in Fig. 2 reflect vertical shear values between 850 and 200 mb that decrease only slightly during the preconditioning phase. Because 6 m s$^{-1}$ is about one-half the threshold shear value that may prevent TC formation (Zehr 1992), one would not expect environmental vertical wind shear between 850 and 200 mb to be a determining (inhibiting) effect for intensification of Flo.

One change during the preconditioning period of Flo is the vertical shear between 300 and 150 mb (Fig. 4b). Although observational coverage is only fair (good) at 150 (200) mb, and 300 mb winds are relatively sparse in the outer region (recall that the vertical shear is calculated between 200- and 600-km radius) of Flo, both the multiquadric and 4DDA analyses have a greater than 35% decrease in 300–150-mb shear during the preconditioning period. It is suggested that this rapid decrease in upper-tropospheric shear is an indication of an upward extension of the TC circulation. This decrease in 300–150-mb shear is indicated schematically in Fig. 2 by an upward extension of the cyclonic wind maximum to about 150 mb, with an approximately equal increase in wind at 850 and 200 mb so that the shear between those two levels remains approximately constant.

Azimuthal-averaged tangential wind changes during this period are illustrated in Fig. 5. Even disregarding the tangential wind analysis within 200–300 km of the center, it is evident that the cyclonic circulation tends to decrease with elevation between 800 and 250 mb at 0000 UTC 13 September (Fig. 5a) Only anticyclonic winds are present above 250 mb at this time. Although the cyclonic winds have evidently not penetrated throughout the troposphere (at least beyond 200–300 km), a warm-core thermal structure may be inferred. Thirty-six hours later (Fig. 5b), the cyclonic tangential winds have increased and appear to have extended upward at outer radii where the analyses should be more reliable. Since both the 850- and 200-mb winds have
increased in the 200–600-km radial band, the vertical shear between these levels appears to have not changed (Fig. 4a). However, the upward extension of the cyclonic winds above 250 mb from Fig. 5a to Fig. 5b is consistent with a decreasing wind shear between 300 and 150 mb in Fig. 4b.

Another perspective of the upper-tropospheric wind changes during the preconditioning phase is given by the 200-mb tangential wind evolution (Fig. 6). Although the 6-h time continuity could be better within 400 km where separate maxima of 4 m s\(^{-1}\) are analyzed at 1200 UTC 13 September and following 1200 UTC 14 September, an outward extension of cyclonic winds to about 1100 km is indicated by more reliable outer wind observations. Notice that the 200-mb cyclonic wind then rapidly contracts to within 600 km of the center by 0600 UTC 15 September, and that the 200-mb winds then become increasingly anticyclonic with time beyond this radius. Since these rapid increases and decreases in cyclonic winds are documented in multiple independent MQ analyses, and in the 4DDA analyses (not shown), they are not considered to be a spurious feature. This "cyclonic wind burst" has not been previously reported to the authors’ knowledge, perhaps owing to the normal paucity of upper-tropospheric observations all around the TC. This cyclonic wind burst is consistent with the upward extension of azimuthally averaged tangential winds from below 250 mb in Fig. 5a to above 250 mb in Fig. 5b.

Evolution of the 200-mb wind fields from the MQ analyses during the preconditioning period leading up to the cyclonic wind burst in Flo is shown in Fig. 7. At 0000 UTC 13 September (Fig. 7a), Flo is a tropical depression at 14°N, 145°E (Fig. 1a) with almost-radial 200-mb outflow turning anticyclonic beyond about 5° latitude from the center. At this time, Ed has an intensity of 45 kt and is near 20°N, 131°E (Fig. 1b). Notice the anticyclone trailing Ed well to the northwest of Flo, and a small upper-tropospheric cyclone to the north with a 20-kt isotach maximum that facilitates outflow from Flo. The major change during the subsequent 36 h leading
The 200-mb eddy flux convergence of relative angular momentum (EFC) based on the MQ analysis (Fig. 8) and the 4DDA (not shown) indicate cyclonic increases of up to 20 m s\(^{-1}\) day\(^{-1}\) during the cyclonic wind burst. These cyclonic EFC values appear to extend inward to about 500 km and outward to 1300 km. However, the EFC values then rapidly decrease to near-zero values around 0600 UTC 15 September, which is when the cyclonic wind burst at 200 mb was decreasing. Although some maxima in the 6-h EFC calculations appear to have an inward propagation as in the Molinari et al. (1995, 1998) studies of Hurricanes Elena and Danny, it is also plausible to interpret the EFC event as being nearly simultaneous over the 500–1300-km radius, which would be consistent with the 200-mb cyclonic wind burst in this radial band (Fig. 6). Rather than a trough–TC interaction, Flo appears to have moved into an environment with an extended band of relative cyclonic motion (Fig. 7b). Whereas some positive EFC values are also calculated at 150 mb (not shown) during 1800 UTC 13 September to 0000 UTC 14 September, it is not possible to track a coherent signal of an EFC maxima toward the center at that level. Thus, the 200-mb cyclonic wind burst appears to be related to the EFC, although not necessarily as a discrete trough propagating inward from large radii as in some previous studies (e.g., Molinari and Vollaro 1989; Molinari et al. 1995, 1998).

The corresponding 200-mb radial wind evolution (Fig. 9) indicates the cyclonic wind burst period is also associated with an enhanced outflow. Azimuthal-average outflows exceed 8 m s\(^{-1}\) between 400 and 1000 km early in the period and later in the cyclonic wind burst values above 6 m s\(^{-1}\) are found between 700 and 1300 km. After 1800 UTC 14 September until late in the life cycle, the 200-mb radial winds tend to be smaller at outer radii except for a temporary maximum of 8 m s\(^{-1}\) near 500 km radius at 1200 UTC 15 September. The vertical structure of the radial outflow during the 200-mb cyclonic wind burst is indicated in Fig. 10. As noted above, azimuthal-average outflows exceeding 8 m s\(^{-1}\) are found between 700 and 1000 km at 200 mb (and 150 mb). Although the details of the distribution within 300 km are not certain, a considerable radial gradient in the 200-mb outflow (Figs. 9 and 10) does exist between the center and the 8 m s\(^{-1}\) isotach. The coarse vertical resolution in these analyses, and lack of observations in the planetary boundary layer, does not represent well the expected large inflow near the surface.

While the outflow maximum during the early portion of the cyclonic wind burst is at 200 mb, the strongest outflow is at 150 mb at 0000 UTC 14 September (not shown) and then again at 200 mb by 1200 UTC 14 September (see second maximum in Fig. 9). These radial wind components are primarily based on the postexperiment cloud-drift winds that were carefully matched with rawinsonde observations by the University of Wisconsin. Merrill and Velden (1996) have analyzed the outflow evolution in time. The radial wind values in Figs. 9 and 10 are generally consistent with their interpretations, which were done in an isentropic coordinate system. Indeed, this consistency with the Merrill and Velden study is considered to be a validation of the MQ and 4DDA analyses in the upper troposphere where the TCM-90 data coverage is considered to be better than at middle levels. Since similar good agreement with...
observations is found in the lower troposphere (Titley 1998), it is presumed that the middle levels are reasonably well represented by the dynamically based analyses.

Another consistent feature in the radial wind structure (Fig. 10) is a strong (2–3 m s⁻¹), deep inflow just below the outflow layer. This compensating inflow in the middle troposphere below the large outflow is reminiscent of the idealized response to an imposed angular momentum flux in a diagnostic study by Holland and Merrill (1984). The time evolution of the azimuthal-average 400-mb radial wind (Fig. 11) has inflow throughout the preconditioning period except for one off-time (0600 UTC 14 Sep) analysis. The horizontal extent and relative maxima of this inflow correspond well with the 200-mb outflow features. Since the 200- and 400 mb analyses are essentially done independently in the MQ, this correspondence is not forced by assumed vertical structure functions. The 4DDA analysis, which has a strong dynamical constraint via the blending with the 6-h model forecast, also has this inflow layer below the outflow maxima during the preconditioning period. The dynamical implications of the inflow would be to spin up the vortex in the midtroposphere.

In summary, both the MQ and 4DDA analyses indicate a cyclonic wind burst in the 400–1200-km azimuthal-average tangential wind (Fig. 12) occurred on 14 September during the preconditioning period prior
to the rapid intensification of Flo. The timing and radial structure of the eddy flux convergence of relative angular momentum are consistent with the tangential wind increases at 200 mb during this period. An increase in the azimuthal-average radial wind outflow at 200 mb (Fig. 9) and inflow at 400 mb (Fig. 11) is contemporaneous with this persistent, large EFC event and the change from anticyclonic to cyclonic winds in this 400–1200-km radial band. Although the net horizontal divergence implied by this outflow at 300-km radius (Figs. 9 and 10) requires that compensating vertical motions must occur somewhere within 300 km of the center, no interior observations are available to determine what that vertical motion distribution might be. Some of the compensating vertical motion probably originates from the inflow layer (Figs. 10 and 11) just below the outflow layer. However, some of the compensating vertical motion within 300-km radius is hypothesized to be subsidence from above. As will be described in section 3c, this subsidence is hypothesized to be the explanation for the warm core to be extended upward (Fig. 3) as in the conceptual model (Fig. 2).

b. Other cases during TCM-90

Similar tangential and radial winds, vertical wind shear, and EFC calculations were also made for Typhoons Ed, Yancy, and Zola during the intensive observing periods of TCM-90. In contrast to Flo, no 200-mb cyclonic wind burst or steady increase in the outer radii anticyclonic winds is detected for Typhoon Ed (not shown). Rather, little evidence of any cyclonic winds greater than 4 m s$^{-1}$ is found above 250 mb after 1200 UTC 14 September when cyclonic 200-mb winds only extend to a radius of 500–650 km. The anticyclonic winds beyond this radius were quasi-steady at 4–8 m s$^{-1}$. Similarly, the azimuthal-average radial wind is also quasi-steady at 4–6 m s$^{-1}$, with no correlation with the intensity changes of Ed. This outflow was detected by the cirrus cloud-drift winds to rise as high as 150 mb, so the absence of a 200-mb cyclonic wind burst to large radii in Ed evidently is not attributable to absence of deep convection. Ambiguous values of EFC from the 4DDA and MQ analyses of Ed make conclusions regarding this potential contribution somewhat tentative. Whereas the 4DDA analysis suggests an energetic, but short lived, 150-mb EFC event at 1200 UTC 14 September that was extended downward to 200 mb by the vertical structure functions, the absence of supporting 200-mb observations for the MQ analysis resulted in a calculation of near-zero EFC values. Consequently, no preconditioning event of the magnitude and character of that for Flo is found for Ed, which also did not experience a rapid intensification.

Titley (1998) also describes two other typhoons (Yancy and Zola) that are included in the 4DDA analyses but not in the MQ analyses. In the case of Typhoon Yancy, a slow change over 54 h from 4 m s$^{-1}$ anticyclonic to 2 m s$^{-1}$ cyclonic in the 400–1200-km radial band, and the absence of a rapid reversal to strengthening anticyclonic winds, could not be interpreted as a cyclonic wind burst. In the case of Zola, the 400–1200-km azimuthal-average tangential wind at 200 mb rapidly increased from $-6$ to $1$ m s$^{-1}$ just 12 h later and then returned to large anticyclonic values over 24 h in conjunction with an EFC forcing event. A second EFC forcing event also led to a change from anticyclonic to about 2 m s$^{-1}$ cyclonic. Although the intensity increased following each of these events, the rate of increase was not as large, or sustained for as long, as in Flo.

c. Interpretation of the preconditioning conceptual model

It is hypothesized that the upward displacement of the warm core in the preconditioning period conceptual model (Fig. 2), and implied in the Flo minus Ed temperature differences (Fig. 3), is created in part by upper-tropospheric subsidence in direct response to the enhanced outflow at 200 mb during the cyclonic wind burst in Flo. An alternate explanation would be the radial outflow triggered more vigorous convection around the center and that indirectly created compensating subsidence and warming inside an “eye.” The thermal structure differences between Flo and Ed (Fig. 3) are consistent with maintaining a more vigorous deep convection near the center. Evidence of higher outflows (150 mb) is available from the geostationary satellite imagery (see also Merrill and Velden 1996).

Whereas the upward displacement of the warm core above 200 mb and the decrease in the vertical shear between 150 and 300 mb (Fig. 4b) are consistent, cause and effect relationships cannot be resolved with this dataset. It is hypothesized (Fig. 2) that the 200-mb cyclonic wind burst in association with the EFC forcing event decreased the vertical shear in the upper tropo-
sphere. Based on thermal wind considerations, the decrease in 150–300-mb tangential wind shear would require a decrease in radial temperature gradient, which might simply be accomplished by displacement of the warm core to higher elevations. From the hypsometric equation, an elevated warm core would imply a deeper sea level pressure, if no compensating cooling effects occur in the intervening layer. While the hypothesized process alone does not guarantee subsequent rapid intensification, it may provide an indirect mechanism to allow rapid intensification in the future.

Based on composites of rawinsondes, Fitzpatrick (1993) found that rapidly intensifying tropical cyclones had much deeper cyclonic flow and, thus, argued that the upper-tropospheric temperatures would be relatively cooler than in less rapidly intensifying cases. This argument is consistent with the cool anomaly in the 300–200-mb layer (Fig. 3) as the cyclonic wind layer was displaced upward (Fig. 5) in Flo in response to the cyclonic wind burst.

Molinari and Vollaro (1990; Fig. 1) detected a decrease in anticyclonic 200-mb wind 48–60 h before deepening occurs. In their case, this cyclonic wind change was due to an approach of a midlatitude trough.

The existence of the enhanced EFC event during the preconditioning period (rather than during the rapid intensification period) seems significant. Inspection of Figs. 13 and 15 of Challa et al. (1998) suggests the eddy fluxes of angular momentum that provide the lateral boundary condition for their numerical simulations have a decreasing trend 24 h prior to the start of the integration. They use the word “catalyst” to imply that the enhanced eddy fluxes may be part of the initiation of the event, but are not essential during later stages (R. Pfeffer 1998, personal communication). This catalyst concept is consistent with an idealized axisymmetric simulation by Challa and Pfeffer (1980) in which only 12 h of angular momentum forcing at the lateral boundary was sufficient to initiate vortex modifications that ultimately led to an intense hurricane. Such a catalyst role may be an appropriate interpretation for the EFC event during the preconditioning period of Flo. At least in this case, rapid intensification followed the EFC event, and the EFC values were then about an order of magnitude smaller during the rapid intensification period. This contrasts with the continual eddy angular momentum flux used to force development of a hurricane-like vortex in some numerical simulations (e.g., Montgomery and Farrell 1993).

4. Rapid intensification stage of Flo

A distinct increase in the intensification rate occurred following 1200 UTC 15 September (Fig. 4). The intensity of Flo increased 23 m s$^{-1}$ in only 18 h, from 44 m s$^{-1}$ at 1800 UTC 15 September to 67 m s$^{-1}$ by 1200 UTC 16 September. The discussion here will concentrate on the changes analyzed in the upper troposphere and beyond 300-km radius, where data coverage is better. Although the first NASA DC-8 mission did provide detailed observations at a flight level of about 195 mb with dropwindsondes to describe some inner-core vertical structures, this mission was centered around 0600 UTC 16 September and, thus, was not until the end of the rapid intensification stage. A second DC-8 mission was centered on 0600 UTC 17 September and provided a dropwindsonde that documented a 891-mb central sea level pressure, which corresponds to a maximum surface wind estimate of 72 m s$^{-1}$ (Fig. 4). Considering the maximum flight-level (195 mb) wind measured by the DC-8 was 56 m s$^{-1}$, the surface wind estimate seems reasonable (if not conservative) and indicates that the rapid intensification period led to a supertyphoon.

The evolution of the 200-mb tangential wind (Fig. 6) in Flo indicates that the cyclonic wind burst had already terminated by 1200 UTC 15 September, so that the azimuthal-average cyclonic winds then extended to about 550 km. At radii beyond about 700 km, the anticyclonic winds increased rapidly, with a brief maximum of 20 m s$^{-1}$ at 1500-km radius at the time of maximum intensity. Although the discrete cyclonic maxima in the inner region at the times of the two DC-8 missions (described above) have validity, the remainder of the analyses within about 300-km radius is not considered reliable.

The primary features of the 200-mb circulation at the time of the first DC-8 mission are shown in Fig. 13. Notice the inner-core cyclonic circulation is confined to about 450-km radius. Several regions of concentrated outflow are illustrated in the MQ analysis. Large cross-contour flow is indicated to the north in a region of good data coverage over the Japanese island network. The strong anticyclonic flow exceeding 30 m s$^{-1}$ about 15° longitude to the east is between a trailing anticyclone and a large upper low farther to the east. Anticyclonic flow is also found at larger radii on the equatorward side. This dual outflow pattern toward the pole and toward the equator is consistent with forecaster rules of thumb (C. Guard, former director, JTWC, 1993, personal communication) for more rapid intensification.

One of the surprising aspects of this outflow pattern is that the azimuthal-average radial wind (Fig. 9) has a relative minimum during the rapid intensification stage relative to the preconditioning phase (or to the subsequent midlatitude interaction stage). Except for a temporary maximum of 8 m s$^{-1}$ near 400-km radius at 1800 UTC 15 September, the average radial winds are about 4 m s$^{-1}$ in the outer region during the rapid intensification. Notice also that the midtropospheric inflow also disappears during this stage (Fig. 11). Merrill and Velden (1996) have previously noted this decrease in average radial motion during the rapid intensification stage of Flo. Thus, a simple secondary circulation model in which the strength of the outflow branch increases in proportion to the central intensity does not apply in this case.
Vertical wind shears between 850 and 200 mb (Fig. 4a) and 300 and 150 mb (Fig. 4b) are even smaller during the rapid intensification stage than in the preconditioning phase. Except for two MQ values of about 6 m s$^{-1}$, all 850–200-mb values are less than 5 m s$^{-1}$. Very consistent 300–150-mb values are found in the MQ and 4DDA analyses, with a trend toward decreasing vertical shear from the beginning of the rapid intensification period following 1200 UTC 15 September to the maximum intensity at 1200 UTC 17 September. The small vertical 850–200-mb wind shears are well below the Zehr (1992) criterion of 12.5 m s$^{-1}$ that are believed to inhibit TC formation. Even though these shear values are below this threshold value, it may not be concluded that smaller shear values are necessarily more favorable for intensification. For example, the mean 850–200-mb vertical shears during Typhoon Ed from the 4DDA analyses was 4.5 m s$^{-1}$ compared to 6.1 m s$^{-1}$ for Flo, and Ed did not experience rapid intensification. Indeed, the vigorous deep convection in the rapidly developing storm may reduce the vertical shear, which may suggest that vertical shear reduction is a symptom, rather than a precursor, of rapid intensification.

As hinted earlier, the maximum in 200-mb EFC is during the preconditioning phase, and then decreases during the rapid intensification stage (Fig. 8). Given the variability in such an eddy calculation, caution is advised in giving too much credence to short-lived EFC maxima. This is especially true near the center; for example, Molinari and Vollaro (1989) suggested that EFC values within 300 km of the center were questionable. However, both the MQ and 4DDA (not shown) EFC evolutions clearly indicate that the angular momentum forcing in the radii beyond 400 km decreases by more than 50% prior to the rapid intensification. At least in this one case of rapid intensification, concurrent external forcing is not a requirement.

A summary of the azimuthal-average tangential and radial winds and the EFC averaged in the 400–1200-km radial band (Fig. 12) confirms these evolutions. A considerable decrease in the radial wind and the EFC after the preconditioning period is again evident. The cyclonic wind increase during the preconditioning period is clearly followed by increasingly anticyclonic winds in this radial band during the rapid intensification stage.

One interpretation of this lack of a direct correlation of secondary circulation enhancement with intensity increases may be that an internal (within 300 km) readjustment is responsible. Perhaps the external forcing during the preconditioning period initiated contraction of an outer rainband of the Hurricane Danny type described by Willoughby (1995, Fig. 2.9), which is also proposed by Molinari et al. (1998). Unfortunately, the near-continuous monitoring with airborne radar available in the Atlantic is not available in western North Pacific. Whereas the DC-8 did have a weather radar, the images were not recorded for later analysis. Although
the science leader (G. Dunnvan 1990, personal communication) on this mission did note concentric eyewalls near the center, the origin or evolution of these features prior to the mission (recall this was near the end of the rapid intensification) is unknown.

Another alternative focuses on the EFC event during the preconditioning triggering radial winds, and the upward extension of cyclonic winds that led to a thermal structure adjustment (Fig. 2). A possible interpretation is that an upward displacement or modification of the upper-tropospheric thermal structure permits a higher maximum potential intensity as described by Emanuel (1988) and Holland (1997).

Unfortunately, the absence of continual internal data in Flo during this critical rapid intensification stage does not allow resolution of the alternate mechanisms. Both external region dropwindsondes and concurrent aircraft radar in the inner region as in the Hurricane Research Division Synoptic Flow Experiments (Franklin et al. 1993) are required during the rapid intensification stage.

5. Rapid decay stage

As indicated in Fig. 4, the intensity of Flo also decreased after having reached maximum intensity at 0600–1200 UTC 17 September, with a decrease of 50 kt in 36 h. Fortunately, the decrease occurred over open ocean prior to Flo striking Japan, or even greater damage would have occurred. In this sense, Flo was a similar case to Hurricane Opal that also filled rapidly before striking the Gulf of Mexico coast.

While the rapid intensity decrease was occurring, the 4DDA analysis 850–200-mb vertical shear values (Fig. 4a) in the 200–600-km radial band oscillated between <4 m s$^{-1}$ (1200 UTC) and <8 m s$^{-1}$ (0600 and 1800 UTC), and the MQ interpolation values remained less than 4.5 m s$^{-1}$. It is assumed here the 1200 and 0000 UTC 4DDA analyses are more reliable than at 0600 and 1800 UTC (off synoptic) as they agree in magnitude with all of the MQ analyses that draw closely to the observations. Thus, vertical shear does not seem to be a primary factor in the rapid intensity decrease. One interpretation is that the intense cyclonic circulation is able to resist the inhibiting effects of relatively weak environmental shear until the intensity and central convection are diminished by other processes.

The interaction of the outflow of Flo with the midlatitude jet at 0600 UTC 16 September (Fig. 13) continued and intensified. At the time (0600 UTC 18 September) of the third NASA DC-8 mission during the period of rapid intensity decrease, the cyclonic outflow was confined to a relatively small radius (Fig. 14). Extreme cross-isobaric flow toward lower 200-mb heights, and thus rapid accelerations downstream, occurred in the outflow toward the pole. Another outflow branch was toward the southeast, part of which curved cyclonically around the large upper low that had nearly overtaken Flo. Another portion of the southern outflow continued southward and joined the tropical easterly jet over the Philippines (not shown). From an evacuation of mass perspective, this outflow pattern would seem to be favorable for intensification, rather than a rapid intensity decrease.

The azimuthal-average anticyclonic winds at radii be-
beyond 900 km decreased during the period of rapid intensity decrease (Fig. 6). As indicated in Fig. 14, the outflow toward the north was more radial without spiraling anticyclonically. Another major change in the outflow pattern from that during the most intense stage (Fig. 13) was the approach of the upper low. Whereas previously the outflow curved anticyclonically as it flowed southward to the west of this upper low, the outflow was now more directly toward and around the upper low (Fig. 14), so that between 10° and 15° longitude to the east the flow was cyclonic relative to the tropical cyclone. Only weakly anticyclonic flow was to the south or to the west of the center beyond 10° latitude radius.

A major change in the 200-mb EFC (Fig. 8) occurs during the midlatitude interaction stage. Whereas the EFC values had been small positive or slightly negative throughout the rapid intensification stage, large positive EFC values appear first at outer radii and seem to propagate inward. However, these eddy angular momentum fluxes are dominated by large (positive) radial components in the outflow jet to the north (Figs. 13 and 14) that are slightly anticyclonic (negative). That is, negative angular momentum is being exported from the cylinder around the cyclone. The EFC magnitudes increase in time as the tropical cyclone approaches the midlatitude jet, not because a midlatitude trough approaches the northern outflow interacting with the tropical cyclone as in the Molinari and Vollaro (1989) case. In the deep Tropics away from a trough interaction, the Coriolis torque would be expected to cause the tropical cyclone outflow to curve anticyclonically into an outflow jet. In this case with a large 200-mb height gradient to the north, the outflow from Flo was a downstream acceleration in an almost radial direction, so that the EFC increased (Fig. 8), but the tangential winds actually became less anticyclonic (Fig. 7). A simple model of large EFC forcing causing an increase in the inner cyclonic winds (intensity) clearly does not apply during this rapid filling stage.

The increase in the azimuthal-average outflow from the rapid intensification stage to the midlatitude interaction stage is evident in Fig. 9 at all radii beyond 300 km. Values exceed 12 m s⁻¹ at the center time of rapid intensity decrease. Here again, the upper branch of the secondary circulation implied by the azimuthal average is counterintuitive, in that the outflow is increasing when the central intensity of the primary circulation is decreasing.

A summary of these 200-mb azimuthal-average (400–1200-km radial band) wind components and EFC trends during the rapid intensity decrease is given on the right side of Fig. 12. After having only small EFC values during the rapid intensification, the rapid climb in EFC forcing during the rapid intensity decrease period is quite striking. The radial wind appears to become increasingly outward concurrent with the EFC forcing. However, the azimuthal-average 200-mb tangential wind in this radial band is essentially constant throughout the decreasing intensity period, rather than increasing in response to the layer outflow.

The vertical and radial structure of the EFC (Fig. 15) at the time of maximum intensity (1200 UTC 17 Sep) provides a clue to the subsequent intensity decrease. The 200-mb EFC maximum found at about 1200-km radius has a limited radial extent that is related to the northern outflow interacting with the midlatitude jet, which was about to begin already at 0600 UTC 16 September (Fig. 13) and was very strong by 0600 UTC 18 September (Fig. 14). The inward displacement of this concentrated 200-mb EFC forcing is also indicated in Fig. 8. However, the special feature in Fig. 15 is that the maximum EFC forcing extends down to 500 mb, which suggests an interaction with a midlatitude jet over a rather large vertical extent.

As in previous figures, a response to this EFC forcing is found in the azimuthal-average radial wind structure 6 h later (Fig. 16). Rather than an inflow as in the 400-mb radial winds during the preconditioning period (Fig. 11), an outflow is initiated in the middle troposphere. Although the maximum values are only 2 m s⁻¹ in the outer regions where the EFC forcing is located, positive (outward) values are found as far inward as 500-km
radius. Although the 400-mb level was perhaps not optimal early in the midlatitude interaction period, maxima of 2 m s⁻¹ occur later during the rapid decay period (Fig. 11).

In summary, the EFC values and radial winds indicate significant midtropospheric interaction with the midlatitude trough to the west and north of Flo while Flo reaches peak intensity. It is hypothesized that the midtropospheric outflow starts to degrade the efficient in–up–out secondary circulation. Clearly the 200-mb outflow is accelerated as the parcels flow down the geopotential height gradient (Fig. 14). However, the deeper layer of EFC forcing (Fig. 15) and the midtropospheric outflow (Fig. 16) are not considered to be favorable. It is proposed that the midtropospheric outflow will be part of a more shallow secondary circulation in which latent heat release will be at lower elevations and at larger radii (Fig. 17). One interpretation is that convectively induced warming at outer radii decreases the radial temperature gradient between the central region and the near environment so that the radial pressure gradient is decreased, and the maximum winds decrease. An alternative interpretation is that such an outer secondary circulation will tend to “rob” some of the deeper secondary circulation that has maximum inflow near the surface and, thus, gains heat and moisture from the ocean to achieve highest θₑ, deepest ascent at the eyewall, and outflow at the highest elevations. As stated above, flow across the 200-mb geopotential gradient accelerates the EFC and outflow at that level. However, the deep EFC that is initiated when the outer circulation of Flo begins to “feel” the midlatitude pressure gradients down to the midtroposphere also creates midtropospheric outflow. As Flo continues northward, the midtropospheric outflow will come closer and closer to the center, and more of the secondary circulation will have ascent over a shallower depth. Because this ascent is at smaller θₑ values, and latent heat will be released over shallower depths and at larger radii, the storm will weaken. In the terminology of Molinari (1998), this interaction with a deeper midlatitude trough would be considered a bad trough.

This hypothesized sequence as the outer tropical cyclone circulation interacts with the midlatitude circulation may explain how the storm can weaken without the central region reaching cold water or passing over land. Neither is it necessary for the inner-core convection to be exposed to the large vertical shear of the midlatitude jet for rapid weakening to occur, as the hypothesized process begins in the outer regions, and increases while the center is still well equatorward of the midlatitude jet.

6. Summary and discussion

a. Data and analyses

This case of rapid intensification and rapid decay of Flo occurred at the same time and in same area as an “ordinary” intensification of Typhoon Ed, which then provides a contrasting case. These storms occurred during TCM-90 and three other field experiments, which provided the best dataset for studying tropical cyclones in the western North Pacific. Having two high resolution in space (50 km, each 50 mb) and time (6 h) analyses of this dataset has allowed more confidence in the interpretations presented here. First, the so-called final TCM-90 analyses were produced with the NCEP mesoscale four-dimensional data assimilation system that used the Eta Model as a basis for continued 6-h assimilation of each intensive observing period during TCM-90. Second, a special set of multiquadric analyses was prepared during the TCM-90 intensive observing period that included Flo and Ed. The MQ technique provides a regional reanalysis that draws closely to observations where available and is bounded to the NOGAPS analysis in data-sparse regions away from the TCM-90 special observations. Another special feature of the MQ analyses is insertion of synthetic observations to better define the TC center location and structure (as can be
resolved on a 50-km grid). Without any synthetic observations, the 4DDA depends primarily on the 6-h model forecasts to provide the tropical cyclone location and inner structure where few observations exist.

Independent flux and budget calculations in storm coordinates with 50-km radial increments have allowed judgments as to the likely reality of the radial variations in these calculations. All of the key features describing the rapid intensity increase and subsequent decrease are present in both analyses. In general, the MQ analyses have been preferred for display because that technique fit the observations more closely. For example, the MQ analysis had larger magnitudes of azimuthal-average tangential and radial winds, and also the derived quantities such as angular momentum fluxes in the upper troposphere. The azimuthal-average radial winds from the MQ analysis also agreed well with the manual, isentropic analyses by Merrill and Velden (1996) of the outflow from Flo.

b. Rapid intensification

A single case study such as this cannot provide necessary and sufficient conditions, or describe the multiple ways in which a tropical cyclone might rapidly intensify. Thus, the purpose here is to summarize some of the special aspects of this case and highlight some results that deviate from other cases or numerical model simulations. This summary is organized in terms of the favorable environmental factors for tropical cyclone intensification in section 1.

Because the SST analyses had values of 28°–28.5°C throughout the region of intensification, the requirement for an SST greater than 26°C is not a factor. Available ocean thermal structure analyses do not indicate warm eddies, such as the Loop Current in the Gulf of Mexico, that might account for rapid intensification. Because of the deep ocean mixed layers in this region, the tropical cyclone-induced SST decreases were generally less the 1.5°C, which did not inhibit Flo from becoming an intense typhoon.

Whereas a maximum potential intensity (MPI) based only on these SSTs would suggest a value of about 70 m s⁻¹, Flo achieved, and probably exceeded, this value and Ed did not. While sufficiently high SSTs are a necessary condition for intense TCs, they are not sufficient to guarantee rapid intensification, or that the TC will eventually reach the MPI. Estimates of MPI by Emanuel (1988) and Holland (1997) also depend on the upper-tropospheric thermal structure. A key comparison of the thermal structures of Flo and Ed at similar intensities indicates a more unstable troposphere and warmer upper troposphere over the central region of Flo prior to the rapid intensification. Thus, a preconditioning stage may have created a more favorable environment for later intensification as proposed by Gray (1992).

Two calculations of vertical shear in a radial band of 200–600 km provide key information. First, the traditional 850–200-mb shear does not appear to be an inhibiting factor to intensification. That is, these shear values are small relative to the Zehr (1992) threshold value. Vertical shears are reduced even further during the rapid intensification. Overall, the vertical shears during Flo were slightly larger than during Ed, which did not rapidly intensify. That is, small 850–200-mb vertical shear may be a necessary, but not a sufficient, condition for intensification.

A special aspect during the preconditioning period of Flo was a 35% decrease in the 300–150-mb vertical shear. This shear decrease occurred in conjunction with a 200-mb cyclonic wind burst that extended to about 1100-km radius. Prior to the rapid intensification, the radius of cyclonic tangential winds reduced to within 600 km of the center, and the outer winds became increasingly anticyclonic until maximum intensity was achieved. Three other typhoons during TCM-90 that did not have sustained rapid intensification were checked, and these did not have such a cyclonic wind burst at 200 mb.

Other variables appeared to have changes consistent with a cyclonic wind burst. A concurrent 200-mb EFC forcing event with magnitudes exceeding 20 m s⁻¹ day⁻¹ would seem to provide an explanation for the cyclonic spinup. These EFC values also decreased to small values concurrently with the end of the cyclonic wind burst. Larger azimuthal-average radial winds also occurred at 200 mb in conjunction with the cyclonic wind burst. The radial variations in this outflow near the center would imply a compensating vertical mass flux. Some of this mass flux was from below in the form of deep convection, with tops extending to 150 mb and higher. It is also hypothesized that some of the compensating vertical mass flux was forced subsidence from the upper troposphere, which then lead to an upward displacement of the warm core mentioned above as a special feature of the thermal structure of Flo relative to Ed. Although how the thermal wind balance is achieved is not clear, a decrease in the vertical shear across 200 mb (i.e., the 300–150-mb shear) is consistent with an upward displacement of the warm core mentioned above. Finally, a layer of inflow was established below the layer of enhanced outflow. Such an inflow is consistent with a further spinup of the cyclonic winds in the middle to upper troposphere.

The observed limited period of EFC forcing during the preconditioning phase of Flo seems to be analogous to the numerical simulations of Challa and Pfeffer (1980) and Challa et al. (1998). That is, the EFC appears to have a “catalyst role,” rather than being required to be a continued forcing as in the Montgomery and Farrell (1993) simulation of tropical cyclone formation.

Unfortunately, many aspects of the rapid intensification of Flo were not resolved because of inadequate aircraft observations in the inner region. Whereas some ideas about environmental conditions during rapid intensification can be confirmed, other ideas cannot be
substantiated. The 850–200-mb vertical wind shear did reach minimum values during the most intense stage of Flo; however, this appears to be more of a symptom rather than a requirement for intensification. The rapid intensification appeared to follow establishment of a second, and strong, outflow channel toward the mid-latitudes, which forecasters have used as a favorable indicator of intensification. However, the azimuthal-average radial wind actually decreased. That is, a conceptual model of the outflow branch of the secondary circulation (in–up–out) being directly correlated with the intensity is not supported. As mentioned above, the EFC values were much smaller during the rapid intensification stage than during the preconditioning phase, so a requirement for the direct EFC forcing is not supported. One interpretation is that once the preconditioning stage had established favorable conditions, an internal process (not resolved by this dataset) resulted in rapid intensification. This could have been an outer rainband contraction such as Willoughby et al. (1982) or Willoughby (1995), which may have been triggered in some way by the EFC event during the preconditioning stage.

c. Rapid decay stage

One case of rapid decay as in this study of Flo cannot be a general description. However, some complex and seemingly inconsistent environmental conditions were observed in this case. First, the rapid decay occurred over open ocean (i.e., not as a result of landfall) with sea surface temperatures that could have sustained a more intense storm. The MQ analyses (or the more reliable 4DDA analyses) of the 850–200-mb vertical wind shears in the 200–600-km radial band did not significantly increase during the rapid decay period. Vigorous upper-tropospheric outflow was observed to the north, with strong accelerations as the flow crossed toward lower pressures. Such an efficient evacuation of mass from the tropical cyclone might have been expected to increase the intensity, rather than occur when the storm was filling. The increasing EFC at 200 mb would also seem to favor intensification. However, the EFC was increased over a layer extending down to 500 mb, presumably as a result of an interaction with a weak mid-latitude trough. This interaction occurred as Flo approached the midlatitudes, and was not a vigorous tropical cyclone–trough interaction as in Molinari and Vollaro (1989) or Molinari et al. (1998). One response to this deeper EFC forcing was the radial wind in the 400–500-mb layer reversed from inflow to outflow. Thus, a shallower secondary circulation that had midtropospheric outflow, and presumably upward vertical motion and latent heat release at lower elevations and at larger radii, is hypothesized to have led to a less efficient energy conversion, and Flo rapidly decayed even though several of the environmental conditions were still favorable.

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