A Three-Dimensional Analysis of the Outflow Layer of Supertyphoon Flo (1990)

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ABSTRACT

Isentropic coordinate analyses of rawinsondes and cloud motion wind vectors derived from geostationary satellite imagery are employed to describe the three-dimensional upper-tropospheric and lower-stratospheric circulation attending western North Pacific Supertyphoon Flo during September 1990. Outflow from the storm is concentrated in several evolving channels in the horizontal. In terms of vertical structure, net outflow evaluated at 6° latitude (666 km) radius is found to occur at higher levels and over an increasing range of potential temperature $\theta$ as the tropical cyclone intensifies. Outflow on the equatorward side of the tropical cyclone tends to occur at greater $\theta$ values (higher altitudes) than poleward outflow. Potential vorticity also decreases within the corresponding isentropic layers associated with the outflow. The implications of the vertical variability of outflow structure in terms of the interactions between storm and environment, and effects on storm structural changes, are considered briefly.

1. Introduction

It has long been contended that tropical cyclone structural changes can be influenced significantly by conditions in the “outflow layer,” typically between 100 and 300 hPa. The mechanism of the influence is still not clearly understood. Emanuel (1986, 1988) proposes a conceptual model of a steady-state tropical cyclone for which the sea surface temperature (SST) and the temperature of the outflow impose a thermodynamic upper bound on intensity. Holland and Merrill (1984), Molinari and Voller (1990), and others suggest that interaction with adjacent weather systems in the outflow layer may alter the secondary circulation of tropical cyclones and stimulate the tropical cyclone’s convection and thereby its intensity. More recently, DeMaria et al. (1993) successfully incorporate upper-tropospheric diagnostics into a statistical model for predicting intensity change. Interest in the outflow layer’s role in structural change has led to descriptive studies (Black and Anthes 1971; Merrill 1988) and numerical simulations (Ooyama 1987; Wu and Emanuel 1994). These treat the outflow layer as a vertically homogeneous entity, and conceptual models of its role in structural change have generally done the same.

The work of Emanuel (1986) on tropical cyclone energetics also has implications in regard to the thermodynamic structure and vertical distribution of outflow. In the boundary layer, surface fluxes increase the saturated equivalent potential temperature $\theta_e$ of inflowing air. Outside the boundary layer, $\theta_e$ is conserved for moist-adiabatic processes so that ascent in convective regions and subsequent outflow occurs approximately along surfaces of constant $\theta_e$. In a tropical cyclone, air ascending in the eyewall should therefore emanate at a higher potential temperature $\theta$ (and therefore higher altitude) than air ascending in the peripheral convection of the storm. It can thus be hypothesized that as a tropical cyclone intensifies its outflow should occur over an increasing range of $\theta$ values as the peak $\theta_e$ characteristic of the eyewall increases. A logical extension of this reasoning is that the actual distribution of outflow over the $\theta$ range should depend upon the radial distribution of convection in the storm. If a particular tropical cyclone is characterized by dominant eyewall convection with little peripheral banding, the outflow should be concentrated near the maximum $\theta$ for a storm of that intensity. These concepts imply that the potential effects of an outflow-level environmental stimulus on the tropical cyclone’s convection (and intensity) may depend on the vertical distribution of the outflow and on the outflow-relative height and depth of the stimulus.

The Emanuel (1986) model also has kinematic implications. Boundary layer drag decreases the absolute angular momentum $M_a$ of inflowing air. Outside the boundary layer in a frictionless axisymmetric vortex $M_a$ is conserved; hence, the flow along $\theta_e$, surfaces must

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also be confined to $M_a$ surfaces. Therefore, eyewall outflow should have the additional property of lower $M_a$ relative to outflow from convective bands at outer radii. However, descriptive studies have shown that the outflow layer is highly asymmetric and is actually the site of the greatest imports of $M_a$ into the storm via eddy fluxes (Holland 1983). Therefore, $M_a$ may not be the best indicator of the kinematic properties of outflow air. Potential vorticity (PV) is an alternative; Schubert and Alworth (1987) indicate that tropical cyclones are an upper-tropospheric sink of PV due to the decrease of heating with height. Outside the convective region, PV is approximately conserved even for individual parcels, and one would therefore expect the outflow from tropical cyclones to have relatively low PV values as the effects of the sink are spread by advection (Wu and Emanuel 1994).

Observational verification of the above hypotheses regarding the $\theta$ and PV distribution in the outflow layer requires observations of wind and temperature at relatively high vertical resolution surrounding a tropical cyclone and extending into the lower stratosphere. Given present remote sensing capability, rawinsondes are required. Furthermore, the tropical cyclone must become sufficiently intense to develop an easily detectable excess in core $\theta$ over that of its environment while traversing the observation network. The case examined in this paper, Supertyphoon Flo (Fig. 1), is suitable because it occurred during an intensive observing period associated with the tropical cyclone motion field experiments conducted in the western North Pacific during August–September 1990 (Elsberry et al. 1990). Section 2 describes the enhanced observational datasets resulting from these experiments and the post-processing strategy employed to evaluate the present hypotheses. Section 3 provides a general description of Flo’s near environment during its evolution as observed in satellite imagery and planar analyses. The level ($\theta$ and pressure) and wind speeds pertaining to features apparently associated with outflow are also traced over time to familiarize the reader with the case. A more quantitative treatment then follows. Section 4 describes the symmetric outflow structure and compares it with the implications of the Emanuel (1986) model. Section 5 presents a summary of the asymmetric structure, specifically the differences between poleward and equatorward outflow, and low- and high-$\theta$ layers within the outflow. These results are then summarized in light of recent research on tropical cyclone structure and interactions with the upper-tropospheric environment, and implications for future research are suggested.

2. Data processing and analysis

Meteorological analysis is usually an undetermined problem, and trade-offs consistent with the analysis goals are always required. For this study, an unbiased estimate of the vertical distribution of tropical cyclone outflow is of utmost importance. The primary hypothesis states that the highest $\theta$ associated with, and peak $\theta$ range of, significant outflow increases with storm intensity. This hypothesis cannot be tested with analyses produced by a data assimilation system in which a numerical weather prediction model is used to provide temporal continuity between analyses unless the model’s depictions of the tropical cyclone’s thermal structure and radial distribution of convective mass flux on the mesoscale are quantitatively accurate. Otherwise, errors in the model would alter the distribution of divergence with height and $\theta$ in the outflow and corrupt the analyses via the first-guess fields. A static, univariate method in which the objective analyses are fitted directly to the respective observations without vertical blending was therefore chosen. It is fully recognized that this choice may result in a lack of (a) temporal continuity, (b) consistency between mass and wind, and (c) “reasonableness” in data voids that a model-based 4D assimilation scheme would attempt to ensure. On the other hand, the presence of some of these properties in a static analysis series would indicate that an image of the true atmospheric state is indeed being obtained.

The datasets employed in this study are those collected during four field experiments conducted in the
western North Pacific Ocean during August–September 1990: 1) TCM-90 (tropical cyclone motion experiment) conducted by the U.S. Office of Naval Research, 2) SPECTRUM (Special Experiment Concerning Typhoon Recurvature and Unusual Motion) conducted by the Economic and Social Commission for Asia and the Pacific/World Meteorological Organization Typhoon Committee, 3) TYPHOON-90 conducted by the Union of Soviet Socialist Republics, and 4) TATEX (Taiwan Area Typhoon Experiment) conducted by Taiwan. A summary of these four field experiments is available in Elsberry et al. (1990). Routine and special observations from all four experiments were assembled by the TCM-90 data management team (Harr et al. 1991). Rawin sondes, dropwin sondes, and reprocessed satellite cloud motion winds (CMWs) from this merged dataset are employed in this study.

In addition to operational CMWs produced by the Japanese Meteorological Agency (JMA), an enhanced set of vectors was postprocessed at 6-h intervals by researchers at the University of Wisconsin—Madison as part of TCM-90 data collection. These high-density vectors were derived from sequential Geostationary Meteorological Satellite (GMS) imagery (IR and VIS at 5-km horizontal resolution) by meticulous manual tracking procedures on a McIDAS (Man–Computer Interactive Data Access System) workstation (Suomi et al. 1983). Subjective quality control was applied by careful cross-validation with nearby postprocessed rawinsonde and dropwinsonde reports, along with spatial and temporal continuity checks. The most difficult aspect of CMW production, altitude assignment, was accomplished using a combination of cloud-top temperatures, comparison with nearby rawinsonde and dropwinsonde reports, and subjective spatial and temporal continuity checks. CMWs provided to TCM-90 were assigned pressure altitudes based on this process. However, for the present study, the winds had to be reassigned to isentropic layers. This was accomplished by comparing each CMW location to the appropriate set of P analyses and using the pressure assigned to the CMW to place it in a θ layer. The winds were then again subjected to consistency checks and compared with rawinsonde averages for each layer. Less than 10% of the wind θ values were adjusted from those originally assigned.

An example of the observational network in terms of typical horizontal coverage during Flo is shown in Fig. 2. In general, the rawinsonde coverage (bold wind vectors in Fig. 2 plotted at 354 K) around Flo during the period of interest was most complete to the north-northwest clockwise through southeast of the storm center and was poorest to the east and northeast. Rawinsonde coverage at 1200 UTC on 16 and 17 September was especially complete. Generally, datasets at 0600 and 1800 UTC contained fewer rawin sondes and were less useful. Omega dropwin sondes (ODWs) at 0600 UTC 17 September and 1200 UTC 18 September provided excellent near-storm coverage below about 250 hPa but were unfortunately not available at higher levels to help define the depth of the outflow. The reprocessed upper-level CMWs provided extensive horizontal coverage in areas of cirrus but are generally confined to a θ range of 350–362 K (thin wind vectors plotted in Fig. 2).

The analysis domain is a pseudo-Mercator grid extending from 2° to 60°N and 95° to 165°E with a 111-km (1° latitude) spacing. Analyses were produced independently for each of 12 θ layers, 4 K thick, beginning with a 338-K layer (336–340 K) and extending through the 382-K layer (380–384 K). The vertical resolution corresponds to about 50 hPa in the troposphere. The analyzed variables are the average pressure (P) of a θ layer, logarithm of pressure thickness of a θ layer (ΔP) and zonal (Wx) and meridional (Wy) components of the average wind in a θ layer. The “recursive filter” analysis method of Hayden and Purser (1995) was used to prepare the analyses. This iterative method features a smooth transition across regions of varying data density, user-specified initial weights for different data types, and adaptive weighting of individual observations based on their consistency with the analysis at each iteration. The filter is initially applied with a scale of 666 km (6° latitude) and smoothly decreases to 37 km (0.33° latitude) during the final pass. For each filter pass, the normalized observation increments from the analysis are computed and observations deviating by more than a user-specified threshold are no longer considered. Increments of Wx and Wy are con-

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Note: θ refers to potential temperature. ΔP is transformed to ln(ΔP) for analysis and back to ΔP for display. This ensures that the analyzed ΔP is always positive.
considered simultaneously so that the same observations are influencing each component during each pass. After several trials the threshold was set to 3 standard deviations for $P$, $W_s$, and $W_r$ and 1.5 standard deviations for $\ln(\Delta P)$. The large values for $P$, $W_s$, and $W_r$ effectively require the analysis to consider nearly all observations at all filter scales, so that all questionable or unrepresentative reports have to be eliminated beforehand by a thorough quality control procedure.

Because of the sensitivity of the chosen analysis method and the limited quantity of reports, subjective quality control was very strict and every effort was made to correct and restore, rather than eliminate, soundings containing errors. Most of the rawinsondes were first subjected to the NOGAPS (Naval Operational Global Atmospheric Prediction System) gross error and internal consistency checks upon receipt in real time at the TCM-90 data collection center at the Naval Postgraduate School in Monterey (Harr et al. 1991). The data were then decoded from the FGGE-IIB [First GARP (Global Atmospheric Research Program) Global Experiment] format distributed by TCM-90 and written to McIDAS (Suomi et al. 1983) compatible files for visual inspection and subsequent processing. All reports flagged as questionable by NOGAPS or late reports including all dropwindsondes not processed through NOGAPS were displayed and inspected. In some cases, obvious encoding errors could be corrected and in other cases the questionable quantity was either deleted or retained based on subjective evaluation of its consistency with the rest of the sounding. In cases where NOGAPS modified a sounding, the original was reconstructed and inspected as well. All soundings in the vicinity of the typhoon were inspected whether flagged with possible errors or not. Visual checks for spatial and temporal consistency were then applied at each mandatory level, and questionable soundings were reinspected. Plots of winds in vector form in most cases highlighted the inconsistent reports. For scalar quantities, a Barnes (1973) analysis was applied using a small influence radius and displayed on McIDAS; inconsistent reports appeared as clear perturbations in the field. The $P$, $\Delta P$, $W_s$, and $W_r$ averages for the isentropic analysis layers were computed for each sounding and evaluated for consistency and continuity for each $\theta$ layer. Because of the anticipated importance of the tropopause location and structure, soundings having only mandatory level temperatures (no significant level or tropopause level reports) were used as part of the quality control on mandatory levels but were not employed in the $\theta$ layer analyses.

Following the quality control, recursive filter analyses of $P$ and $\Delta P$ for all layers and analysis times were prepared as an initial step. These served as the basis for the final processing (height assignment) of the CMWs. Wind analyses (zonal $W_r$ and meridional $W_r$ components) were then prepared for each $\theta$ layer— one set containing the reprocessed CMWs and one without. Several methods of introducing the CMWs were attempted, including both successive and simultaneous analysis of rawinsondes and CMWs. Simultaneous analysis with CMWs given half the weight of rawinsondes was eventually chosen empirically to allow the CMWs to control the analysis between rawinsondes yet not overwhelm rawinsondes when the latter were present. A limitation of CMWs for this application is that even if assigned to the correct $\theta$ layer, they are single-level measurements and in some cases may lead to aliasing of features in the vertical.

Additional diagnostic quantities were then derived from the analyses of $P$, $\Delta P$, $W_s$, and $W_r$. In addition to the pseudo-Mercator analysis grid, a cylindrical coordinate system moving with the storm was also employed. In this system, radial wind ($u$) is positive outward and tangential wind ($v$) is cyclonic for a Northern Hemisphere storm (Typhoon Flo). From analyses of $P$, $Z$, $\Delta P$, $W_s$, and $W_r$ for isentropic layers, $u$ and $v$ were computed for each grid point, absolute vorticity ($\zeta$) was computed for all interior points by finite differencing of the $W_r$ (zonal wind) and $W_r$ (meridional wind) fields and addition of the locally computed Coriolis term, and PV was computed from $\Delta P$ and $\zeta$. Constant radius averages of these quantities were then computed at radii ranging from 111 to 1666 km (1° to 15° latitude) for each $\theta$ layer to produce $R-\theta$ sections. The [$\Box$] operator is used to denote a constant radius mean spanning a selected sector; if the subscript is omitted, an "axisymmetric" mean is implied.

As a further aid to interpretation, the Lagrangian coordinate (storm relative) datasets were also converted from $\theta$ to height ($Z$) coordinates for multidimensional display using VIS-5d. The VIS-5d package (Hibbard et al. 1994) allows interactive visualization of multivariable datasets in space and time in the form of plan views, cross sections, material surfaces, or volume rendering of each variable. The analyses are converted to $Z$ coordinates by analyzing $Z$ on $\theta$ surfaces using the recursive filter analysis described above. Values of each variable at fixed increments in $Z$ (0.5 km from 7 to 17 km) were then linearly interpolated from analyzed values at the $\theta$ surfaces nearest in $Z$.

3. General overview of the upper-level wind environment and evolution

To familiarize the reader with the case before beginning a more quantitative analysis of Flo's outflow layer, a general qualitative description of the upper-level environmental wind conditions around Flo during the 72-h period of interest is presented in this section. The description is based on a series of GMS infrared images (Fig. 3) at 12-h intervals, accompanied by text indicating the location and heights ($\theta$ and pressure) of the relevant cloud and circulation features as determined from observations and the analyses. This description emphasizes synoptic patterns, satellite signatures, and
FIG. 3. Infrared window channel imagery from the GMS, remapped into a Mercator projection with 10-km resolution: (a) 1200 UTC 15 September 1990, (b) 0000 UTC 16 September, (c) 1200 UTC 16 September, (d) 0000 UTC 17 September, (e) 1200 UTC 17 September, (f) 0000 UTC 18 September, and (g) 1200 UTC 18 September. The letter symbols in (a) introduce synoptic features of interest and are described in the text.
data distribution. Later descriptions (sections 4 and 5) emphasize the quantitative symmetric and asymmetric structure of the outflow.

a. 15 September 1990

At 1200 UTC 15 September, Flo (Fig. 3a, feature A) is a steadily deepening minimal typhoon (MSW of 40 m s$^{-1}$) of average size, with a prominent convective band to the north and northeast. Flo’s track and intensity during the 72-h case study period is shown in Fig. 1. Typhoon Ed (feature B), also a minimal typhoon, moves generally westward away from Flo during the 72-h period and reaches a peak intensity of 45 m s$^{-1}$ on 17 September. A tropical upper-tropospheric trough (TUTT) cell is located to the east (feature C) and moves steadily west-northwestward and closer to Flo, during the period. A rawinsonde observation indicated that the TUTT was cold cored up to about 200 hPa (350 K) and warm cored above. The cloud band along 30$^\circ$N and extending east-northeast over Japan (feature D) is associated with an upper-tropospheric baroclinic zone and belt of westerly winds above 400 hPa. The lower-tropospheric baroclinic zone associated with the polar front is located 15$^\circ$—20$^\circ$ latitude farther to the north. These two streams of westerlies and Flo eventually merge over Japan toward the end of the case study period.

Figure 3a reveals that an extensive cirrus shield tracing anticyclonic flow to the north and east of Flo is separated from the cloud band to the north associated with the westerlies. CMWs indicate an area of 15—35 m s$^{-1}$ northwesterly winds on the outer edge of the cirrus shield, well to the east-northeast of Flo. Comparisons of these CMWs with nearly collocated rawinsondes place this cirrus near 350 K (180 hPa). High-resolution animated imagery (Velden 1990) implies that it originated from Flo about 2 days earlier. The objective analyses suggest that the strongest outflow, in terms of storm-relative radial wind $u$, is occurring at 354 K (about 160 hPa) in the northeast quadrant. Animated satellite imagery indicates another region of apparent outflow, a thin anticyclonically rotating cirrus layer within 600 km of the typhoon center in the southeast and south quadrants. Unlike the persistent cirrus to the north and east, it appears to dissipate relatively quickly as it disperses outward. The few CMWs associated with this cirrus indicate anticyclonic flow of 10—20 m s$^{-1}$ at a level near 358 K (140 hPa). This stream appears to split, with one branch curving anticyclonically around the west side of Flo and the other flowing equatorward and confluent with outflow from Typhoon Ed.

b. 16 September 1990

Twelve hours later, at 0000 UTC 16 September, Flo has begun rapid deepening (MSW has increased to 52 m s$^{-1}$). The satellite presentation (Fig. 3b) indicates an expanding central dense overcast (CDO) and a larger, clearer eye. Expanding cirrus outflow to the west of Flo is estimated to be near 354 K (170 hPa) with east-southeasterly winds exceeding 10 m s$^{-1}$. An increase in somewhat higher level (358 K and 140 hPa) cirrus and outflow is also noted to the east and southeast of Flo.

By 1200 UTC, Flo has attained a "supertyphoon" intensity of 67 m s$^{-1}$ and continues to deepen. The CDO has continued to increase in size, and a distinct eye is evident (Fig. 3c). CMWs along the edge of the CDO and within the surrounding cirrus suggest outflow in all quadrants. Rawinsonde reports and expanding cirrus imply that outflow extends 1000 km to the west of the center at 354 K (180 hPa). This outflow still remains distinct from the midlatitude cloud band and associated westerlies at this time. The peripheral convective band noted north of Flo’s CDO at 1200 UTC 15 September has detached and is now situated between Flo and the midlatitude cloud band over Korea. The latter has also become thicker in the past 24 h and a vigorous short-wave trough ("comma tail" over Manchuria) is approaching from the west-northwest. To the east and south of the center, rawinsondes and CMWs show that the outflow has increased in areal extent to 800-km radius, strengthened to over 15 m s$^{-1}$, and has increased in altitude ($\theta$) to 362—366 K (120—130 hPa). An equatorward flow maximum of 35 m s$^{-1}$ at 354 K (160 hPa) 1500 km east of Flo is still evident between Flo and the TUTT cell, although the cirrus there has eroded considerably compared with a day earlier.

c. 17 September 1990

As Flo achieves its peak intensity of 75 m s$^{-1}$ and begins to recurve, the cloud signature evolves from zonally oriented (most pronounced in Fig. 3b at 1200 UTC 16 September) to meridionally elongated (Fig. 3e at 0000 UTC 18 September). Changes also occur in the relative strengths of surrounding features. As the short wave previously over Manchuria crosses into the Sea of Japan, the cloud band associated with the upper-tropospheric baroclinic zone over China and the outflow cirrus west of Flo largely dissipate, leaving a relatively clear region west of 125$^\circ$E. Convection in the former north-side peripheral band strengthens briefly around 0000 UTC 17 September (Fig. 3d) before this feature, the shortwave, the zonal cloud band, and a new plume of Flo outflow cirrus merge over Japan and the Sea of Japan by 1200 UTC (Fig. 3e). Outflow to the northwest near 354 K (180 hPa) is now confluent with the midlatitude southwesterlies over the southern Korean peninsula. Outflow to the south and east near 130 hPa (358—366 K) is still evident from thin cirrus tracers out to 1100 km from the center. Backing and strengthening of a rawinsonde wind profile at 20$^\circ$N,
126°E at $\theta$ values up to 382 K between 1200 UTC 16 September and 0000 UTC 17 September indicates that this outflow has become very deep. The col between Ed and Flo has shifted southward to northern Luzon as Ed moves off to the west, and the easterly outflow south of Flo now extends westward to the Philippines.

d. 18 September 1990

Flo begins to weaken steadily on 18 September with maximum winds diminishing to 57 m s$^{-1}$ by 1200 UTC. A concentric eye and a warmer, less symmetric CDO are evident (Figs. 3f–g). Willoughby (1990) and Dvorak (1984) relate these features, respectively, to weakening. Rawinsonde also indicate that cooler, drier air is entering the circulation from the northwest. The previous trend toward meridional elongation in the satellite presentation continues, and an expansion of thicker cirrus in the equatorward outflow south of Flo is also evident for the first time.

At 0000 UTC, CMWs indicate wind speeds along the cirrus band edge northwest of Flo range from 15 m s$^{-1}$ 600 km due west of Flo to 50 m s$^{-1}$ 1100 km to the northwest. Rawinsondes indicate that the peak outflow is near 354 K (180 hPa). To the south and east, outflow is found at 358–382 K. By 1200 UTC outflow from the south side is evident at two levels. A lower (dominant) stream near 170 hPa (354 K) proceeds southward and southeastward to 1000-km radius and beyond, confluent with the strong northerlies to the east between Flo and the nearby TUTT cell. The higher stream near 130 hPa (358–370 K) appears to quickly turn anticyclonically around the west side of Flo within 500 km of the center and overlaps the poleward outflow emanating from lower altitudes.

Several features evident from this preliminary discussion are the following.

1) The overall cloud (and outflow) pattern evolves from zonally oriented to meridionally elongated as adjacent weather systems enter Flo’s environment.

2) The $\theta$ values and pressure altitudes of the outflow tend to increase during Flo’s deepening stage.

3) When compared with poleward outflow, equatorward outflow is generally higher (in terms of both $\theta$ and pressure) and is characterized by thinner, short-lived cirrus.

In the following sections, the development of the outflow and its properties will be examined more quantitatively in terms of symmetrically averaged quantities, means for north and south semicircles relative to the storm, and analyses of total and radial wind in two representative $\theta$ layers.

4. Vertical and temporal variability of storm-relative symmetric flow

The primary hypotheses of the paper regarding changes in the $\theta$ range (depth) of the outflow and overall distribution of PV with respect to $\theta$ surfaces can be addressed most directly using axisymmetric averages. Radius–potential temperature sections were prepared from the rawinsonde-plus-CMW analyses according to the method described at the end of section 2. Figure 4 shows the axisymmetric structure of the outflow at 1200 UTC 17 September. This is during peak intensity; it also corresponds to the time of overall best data coverage. Figures 5–8 then highlight the temporal changes in the distributions of various quantities. Axisymmetric means must be interpreted cautiously, especially at large radii where asymmetries are large and environmental features may predominate.

Figure 4a shows the thermal structure; the warm core is evident as an inward increase in $P$ in the $\theta$ layers at or below 358 K (160 hPa); a weak cold core is evident at larger $\theta$. The top of the warm core is closely aligned with the tropopause, evident as a layer of decreasing pressure thickness ($\Delta P$) from 358 to 362 K (130 to 170 hPa) near the center and 350 to 354 K (190 to 240 hPa) at outer radii (Fig. 4b). The tropopause is therefore deformed both to lower $P$ and higher $\theta$ over the storm. The development of the warm core mirrors the intensity tendency. During the period of rapid deepening (1200 UTC 15 September through 1200 UTC 16 September) $P$ increases by up to 30 hPa on $\theta$ surfaces within 500 km of the center in the 350–354-K range (Fig. 5), and the tropopause $\theta$ indicated by the $\Delta P$ field increases from about 350 K to around 358 K by 1200 UTC 17 September near the storm center. After 0000 UTC 18 September (not shown), the $P$ of $\theta$ surfaces at $R > 800$ km increases above 350 K and decreases below as the storm moves closer to the relatively cold upper troposphere and warmer lower stratosphere of the baroclinic westerlies.

Figures 4c–f show various aspects of the kinematic structure of Flo at peak intensity. Significant outflow is occurring inside 1000-km radius between about 110 and 240 hPa or 350 and 374 K. As noted in the introduction, a primary purpose of this research is to examine the variability in the height and depth of the outflow over the life cycle of the storm. Figure 6 shows time–potential temperature sections at a radius of 666 km (6° latitude) computed from both rawinsonde-only and rawinsonde-plus-CMW analyses. This radius is chosen because it is approximately the radius of maximum radial wind as resolved in these analyses. Qualitatively similar behavior is observed at 8° latitude. The choice of a particular threshold [$u$] to define the outflow layer is somewhat arbitrary, but the value chosen should be large enough that it is exceeded only when a significant feature (such as a tropical cyclone) is present within the prescribed radius. This consideration was applied by recomputing $u$ and then [$u$] at 666-km radius relative to each point on the analysis grid (instead of just the tropical cyclone center) and displaying the resulting contour maps (not shown) of [$u$] relative to each grid point (essentially an area-averaged diver-
Fig. 4. Axisymmetric means of analyzed (rawinsonde plus CMW) variables in a cylindrical coordinate system centered on Typhoon Flo at 1200 UTC 17 September: (a) $P$ (hPa), (b) $\Delta P$ (hPa (K$^{-1}$)), (c) $u$ (m s$^{-1}$), (d) $v$ (m s$^{-1}$), (e) $\zeta$ (10$^{-5}$ s$^{-1}$), and (f) [PV] (0.1 PVU).

Large areas of $[u]$ with magnitudes of less than 2 m s$^{-1}$ were observed, but local maxima exceeding 2 m s$^{-1}$ were generally associated with identifiable weather systems (including Flo) and showed temporal continuity. Therefore, $[u] > 2$ m s$^{-1}$ was adopted as the threshold identifier of "significant" outflow at 666-km radius. This corresponds to an average divergence of $6 \times 10^{-6}$ s$^{-1}$ over the region bounded by 666-km radius.

The overall pattern depicted in Fig. 6 is in general agreement with the qualitative observations outlined in
section 3. Addition of the CMWs (Fig. 6b) mainly affects the trend analysis of the bottom of the outflow layer. The lowest \(\theta\) value at which significant outflow is occurring never rises above 350 K (240 hPa), and the peak values are generally stronger. The primary effect of the CMWs is to fill in some of the rainsonde data voids, especially prominent on the east side of the storm where the CMWs are concentrated in the 350–358 K layer. The CMW levels (\(\theta\) values) are assigned based on comparisons with nearby rainsondes in space and time and then “propagated” into rainsonde-free areas by maintaining continuity within fleets of CMWs associated with cirrus of similar appearance and motion. The possibility exists that the CMWs are assigned potential temperatures that are systematically too low and that the bottom of the outflow has indeed risen to higher \(\theta\) during the period 1200 UTC 16 through 1200 UTC 17 September (Fig. 6a), although this is not actually believed to be the case. It is more likely that the CMWs provide a better definition of an otherwise data-void area to the east where indications are that the outflow is prominent at these lower altitudes.

Outflow is initially observed in a relatively narrow \(\theta\) range (Fig. 6). As the storm intensifies, the \(\theta\) range over which outflow is occurring increases sharply. Late in the period, the \(\theta\) range of outflow decreases as the storm weakens. Figure 7 summarizes the temporal changes of the 2 m s\(^{-1}\) \([u]\) isotach and storm intensity. From Figs. 6b and 7 it is evident that prior to 1200 UTC 16 September the outflow layer is relatively shallow and extends from 350 to 358 K (160 to 230 hPa). From 0000 UTC 16 September through 0000 UTC 17 September the \(\theta\) at the top of the outflow layer increases sharply to about 377 K (110 hPa). By 0000 UTC 18 September the top of the outflow layer peaks near 382 K and then descends to around 370 K (120 hPa) by 1200 UTC 18 September.

It is interesting from Fig. 7 that the top of the outflow layer closely follows the storm intensity with a time lag of one analysis period, or 12 h. This time lag is reasonable given the fact we are analyzing the outflow characteristics at a 666-km radius from the storm center. The analyses do not resolve the strong winds near the core and are too far apart in time for computing precise trajectories. Therefore, a parcel’s origin (eye-wall or peripheral convective band) can be subjectively implied from visual inspection of animated satellite imagery but cannot be specifically determined. A strong positive radial wind at a given point or location upstream from the eyewall region on a streamline does not alone assure an eyewall source for a parcel. This is especially true if the distance from the storm is large. However, a simple calculation can reveal reasonable bounds on the distance traveled by parcels emanating from the eyewall region. A typical \([u]\) of 4 m s\(^{-1}\) at 800-km radius implies a mean divergence over the enclosed area of 10\(^{-5}\) s\(^{-1}\). If the actual divergence is all occurring within 200 km of the center, \([u]\) at 200 km is 16 m s\(^{-1}\), and \([u]\) outside 200 km varies as the inverse of the radius. Integration yields a travel time of around 1 day from the edge of the divergent region to 800-km radius. Concentration of the outflow into channels (as observed in this case) would further decrease the time required for eyewall outflow to be felt at outer radii. Therefore, at 666 km and during peak intensity (strong, channeled outflow), our subjective estimate of parcel travel time is 12–18 h, in general agreement with the intensity versus outflow-top trends shown in Fig. 7.

An independent estimate of the \(\theta\) range of outflow from a storm of Flo’s intensity may be obtained from Emanuel’s (1986) model. The goal is not to validate Emanuel (1986) model’s prediction of a peak intensity for given environmental conditions but rather to diagnose the expected thermodynamic properties of the outflow of an approximate steady-state analog to Flo at peak intensity. The model is relatively sensitive to the SST, outflow temperature, and ambient relative humidity; these factors were adjusted slightly from observed or climatological western North Pacific values until Flo’s observed intensity was matched. The model parameters shown in Table 1 are all well within reasonable ranges of climatological values to the extent known.

For an approximate steady-state analog to Flo at peak intensity, the Emanuel (1986) model predicts a \(\theta_e\) of 380 K at the radius of maximum wind (RMW), with an outward radial \(\theta_e\) gradient of \(-0.60\) K km\(^{-1}\). For ascent in the eyewall at 0–5 km inside the RMW, updraft \(\theta_e\) of 380–383 K and outflow at corresponding \(\theta\) is expected. Outflow emanating at lower \(\theta\) should have its source in convective bands on the periphery of the storm. Undulate ascent of environmental boundary-
layer air would result in outflow near 362 K. However, Powell (1990, Fig. 14) indicates a midtropospheric \( \theta_e \) in an Atlantic hurricane rainband that is 8 K less than that of the environmental boundary layer. Assuming a similar decrease in Flo yields an estimated \( \theta_e \) of 354 K for outflow produced by peripheral convective bands. Outflow at intermediate \( \theta_e \) values between 354 and 380 K should emanate from convection between the eyewall and the periphery. The observed \( \theta_e \) range of outflow emanating from Flo is in general agreement with that diagnosed using the Emanuel model.

The strongest mean anticyclonic flow \( [v] \) < 0 is outward of the \( [u] \) maximum at 1300 km radius and is located in the 358–370-K (110–160 hPa) layer as shown in Fig. 4d. The peak value of 20 m s\(^{-1}\) is the strongest observed during the analysis period. The anticyclone strengthens prior to peak storm intensity at 1200 UTC 17 September with almost all of the anticyclonic tendency found above 350 K. This is consistent with the development of the warm core; the strongest region of anticyclonic vertical shear at 1200 UTC 17 September is found between 346 and 358 K (160 and 300 hPa) inside 1000-km radius and is located just below the tropopause and within the warm core.

The fitted Emanuel (1986) model also predicts tangential winds in the outflow layer based on the assumption that \( M_e \) and \( \theta_e \) surfaces are coincident above the boundary layer. The observed \([v]\) at 700-km radius of \(-7\) and \(-9\) m s\(^{-1}\) at 354 and 380 K, respectively, are cyclonic relative to the predicted \(-13\) and \(-17\) m s\(^{-1}\) of peripheral and eyewall outflow, respectively. Such a difference is to be expected because tropical

<table>
<thead>
<tr>
<th>Specified</th>
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<tr>
<td>Sea surface temperature</td>
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</tr>
<tr>
<td>Air-sea temperature difference</td>
<td>77.5%</td>
</tr>
<tr>
<td>Ambient relative humidity</td>
<td>1000 hPa</td>
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<tr>
<td>Ambient pressure</td>
<td>500 km</td>
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<tr>
<td>Storm radius</td>
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</tbody>
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<table>
<thead>
<tr>
<th>Diagnosed</th>
<th>898 hPa</th>
<th>74 m s(^{-1})</th>
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<tr>
<td>MSLP</td>
<td></td>
<td>39 km</td>
</tr>
<tr>
<td>Ambient boundary layer ( \theta_e )</td>
<td>362 K</td>
<td></td>
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<tr>
<td>Boundary layer ( \theta_e ) at RMW</td>
<td>380 K</td>
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<tr>
<td>Eyewall (380 K) outflow tangential wind at ( R = 666 ) km</td>
<td>(-17) m s(^{-1})</td>
<td></td>
</tr>
<tr>
<td>Peripheral (364 K) outflow tangential wind at ( R = 666 ) km</td>
<td>(-13) m s(^{-1})</td>
<td></td>
</tr>
</tbody>
</table>
cyclone outflow is actually highly asymmetric, resulting in large eddy imports of $M_{\omega}$. The eddy flux convergence of relative angular momentum

$$\text{EFC} = -\frac{1}{r^2} \frac{\partial}{\partial r} (r^2 [u'v'])$$

was evaluated for the $R-\theta$ sections for each time period. Values of less than 5 m s$^{-1}$ per day are observed through the rapid deepening period (1200 UTC 15 September through 1200 UTC 16 September). Environmental assistance to intensification via positive EFC and stimulation of the mean secondary circulation as noted in other cases by Molinari and Vollaro (1990) apparently was not a factor in Flo's rapid development. De Maria et al. (1993) note that such effects are not essential for deepening; many intensifying Atlantic storms are associated with EFC of less than 10 m s$^{-1}$ per day. A rawinsonde composite analysis of deepening western North Pacific storms yields a similar value of around 4 m s$^{-1}$ per day (Holland 1983). As the distance between Flo and adjacent environmental features (the midlatitude westerlies and the TUTT cell discussed in section 3) diminished on 17 September, the EFC at outer radii (10º-15º latitude) was rising to 15-20 m s$^{-1}$ per day by the time of peak intensity (1200 UTC 17 September) and eventually rose to 30-40 m s$^{-1}$ per day at closer radii by 1200 UTC 18 September. The peak values are similar to those found by Molinari and Vollaro (1990) for a single case and De Maria et al. (1993) for a multiyear sample of Atlantic hurricanes experiencing strong environmental effects. The strongest $M_{\omega}$ imports occurred in the north semicircle in the 346-366-K layer. At the same time, the strength of the mean outflow increased (Fig. 6) in the same layer, implying that the stimulation of the secondary circulation by environmentally induced EFC as described by Molinari and Vollaro (1990) may now have been taking place but mainly at levels below those associated with eyewall outflow and only after Flo had already reached peak intensity and was weakening in terms of central pressure (discussed further in section 6).

Absolute vorticity $\zeta$ and PV require computation of horizontal gradients of the wind fields and therefore tax the dataset even further. To compute the axisymmetric $\zeta$ or PV adequately for an annulus about the storm, there should ideally be a double ring of observations straddling it. Only at 1200 UTC on 16 and 17 September is this condition met for the rawinsonde-only set. Nevertheless, definite trends are evident in the $R-\theta$ sections at other times (not shown), giving additional confidence that the overall signatures are valid. As the storm intensifies, the minimum in $\zeta$ moves inward and toward higher $\theta$, beginning at 1100–1400 km and 350 K (180–240 hPa) at 1200 UTC 15 September and being found inside 600 km between 358 and 370 K (120–180 hPa) by 0000 UTC 18 September. The axisymmetric $\zeta$ field at 1200 UTC 17 September (Fig. 4e) shows a minimum coincident with the outflow maximum and sloping toward lower $\theta$ with increasing $R$ out to about 1000 km. Beyond that, an increase with radius with only weak variation in $\theta$ is observed. The value of the Coriolis parameter $f$ at the storm latitude is $6.4 \times 10^{-7}$ s$^{-1}$, indicating that nearly the entire 15º latitude radius domain is occupied by anticyclonic relative vorticity in an axisymmetric sense.

The distribution of PV at 1200 UTC 17 September (Fig. 4f) is dominated by the variations in $\Delta P$ with generally low values in $\theta$ layers 350–358 K and below. Because the $\theta$ of the "static tropopause" (region of $\Delta P$ decrease) lowers with increasing radius, there is a corresponding increase in PV with radius in a given $\theta$ layer in the above range due to the $\Delta P$ distribution alone. The wind field increases this effect; the general increase in $\zeta$ with radius outside about 500 km means that the "dynamic tropopause" (typically considered to be 1.5 PV units (PVU) or $1.5 \times 10^{-6}$ m$^2$ kg$^{-1}$ s$^{-1}$ K) slopes even more sharply relative to $\theta$ surfaces than the static tropopause.

Comparison of $|u|$ with $|\Delta P|$ and [PV] indicates that significant outflow is occurring above the environmental tropopause, at least in an axisymmetric mean sense. Further evidence is provided by the temporal changes of [PV]. As explained by Hoskins et al. (1985) and Haynes and McIntyre (1987) the effect of heating due to condensation is to produce a PV source below the heating and a sink above. Figure 8 shows the time-averaged PV in the $R-\theta$ plane from 1200 UTC 15 September through 1200 UTC 17 September (period of deepening), and the 48-h change as a percent of this average. In the 700–1000-km belt, PV diminishes with time between $\theta$ of 350–380 K, which is associated with advection of source PV within the layer.
of significant outflow \( ([u] > 2 \text{ m s}^{-1}) \). The strongest normalized tendency \( \Delta PV/PV = -1.0 \) is occurring in the 700–900-km band near a \( \theta \) of 362 K, coincident with the \([u]\) maximum as resolvable in this dataset. The greatest absolute decrease from 3.0 to 1.5 PVU is centered on 1000-km radius and 366 K. Tendencies over individual 12-h periods also indicate an upward shift of the normalized PV tendency minimum with time. For the 12 h ending at 0000 UTC 16 September, the top of the PV minimum was 362 K, and by 1200 UTC 17 September the top of the minimum had risen to 376 K. These [PV] signatures are surprisingly consistent, given the data limitations mentioned above.

5. Storm-relative asymmetric flow aspects

Having established a positive relationship between outflow \( \theta \) range and storm intensity in the symmetric sense, it is now of interest to examine the asymmetric structure. As described in section 3, the outflow distribution evolves considerably during this period of Flo’s life cycle. Prior to rapid deepening, outflow was primarily concentrated to the north through east (Fig. 3a). As Flo deepened and reached peak intensity (Figs. 3b–d), the cloud and outflow pattern becomes more symmetric. Finally, the cloud pattern becomes elongated meridionally during the early dissipating stage due to apparent environmental influences (Figs. 3e–g).

Maps of vector and radial wind in two representative \( \theta \) layers—354 and 366 K (Figs. 9 and 10) and \( R-\theta \) sections of radial wind partitioned into north and south semicircles around the storm (Fig. 11) are shown to illustrate the systematic differences in the horizontal and vertical distribution of outflow. The 354-K layer represents the \( \theta \) range of peak \([u]\) throughout the analysis period (Fig. 6b). It will be referred to as “low outflow.” The 366-K layer is chosen because it is the location of a relative maximum in \([u]\) at peak intensity.
and is therefore easily identifiable, and $[u]$ is nearly zero there prior to rapid deepening. It is also qualitatively similar in overall appearance to the outflow occurring at $\theta$ values of 378 K, near the top of the range. The 366-K layer will be referred to as "high outflow." The 366-K layer is also at or above the axisymmetric mean dynamic tropopause (1.5 PVU) throughout the analysis period. Plan views of PV in the lower stratosphere are relatively noisy and are therefore not shown. The PV is very sensitive to small variations in the temperature sounding when $\Delta P$ becomes small, as is the case for 4-K $\theta$ layers in the stratosphere, and the lack of CMWs above 362 K prevents a detailed analysis of the potential vorticity field on the scale of the outflow plumes.

Caution must be exercised when interpreting flow fields in terms of outflow from Supertyphoon Flo. In the following discussion, particular attention is paid to the 600–1000-km radius band, since inside this region the analysis detail is questionable and outside it the distances (and parcel travel times) become large. The $R-\theta$ sections for the storm-relative north and south semicircles also require careful interpretation, as they may be sensitive to features other than Flo’s circulation. The $[u]$ sections shown in Fig. 11 do not just isolate on the typhoon outflow. Especially at larger radii, environmental features can also affect the profiles. For example, the strong band of westerlies to the north of Flo results in large areas of $u > 0$ and $u < 0$; $[u]_{boush}$ constitutes the average of these and therefore may be unrepresentative of the actual speed at which air is flowing outward from the storm. The $[u]_{koush}$ analyses are influenced at large radii by the outflow from Typhoon Ed early in the period and by the TUTT circulation at later time periods.

At 1200 UTC 15 September Flo is a minimal typhoon at the onset of rapid deepening. Figure 9a indicates diffluent southeasterly flow across the storm at 354 K. One branch curves sharply anticyclonically and merges with an area of 10–15 m s$^{-1}$ northerlies between Flo and a TUTT cell (see Fig. 3a). The other branch gently curves cyclonically and merges with another area of northerlies (outflow from Typhoon Ed) near the Philippines. The axisymmetric mean radial

Fig. 10. As in Fig. 9 except for the 366-K $\theta$ layer.
wind \( u \) at 700-km radius is about 5 m s\(^{-1} \), with a radial wind maximum in the north through east and a secondary maximum in the southwest. Figure 10a indicates that Flo lies just south of an eastward extension of a large-scale ridge and appears not to have yet influenced the flow in this layer (366 K) since the \( u \) at 700-km radius is near zero. In the semicircle means \( u_{\text{north}} \), and \( u_{\text{south}} \) (Fig. 11a), systematic differences are already evident. Peak \( u_{\text{north}} \) is indicated near 352 K at 500–600-km radius. The \( u_{\text{south}} \) outflow maximum is found at a much larger radius near \( \theta \) values of 354–356 K. As mentioned earlier, at large radii from Flo it becomes less clear if the mean radial winds are representing pure "outflow." Inspection of Fig. 9a indicates a region of strong northerlies over and to the east of the Philippines, likely associated with Typhoon Ed. This flow contributes to a strong \( u_{\text{north}} \) but does not appear to be directly associated with Flo. In general, the equatorward outflow pattern is initially a weak, relatively thin layer found at slightly higher altitudes than its poleward counterpart.

By 1200 UTC 16 September, Flo is a supertyphoon having just undergone rapid deepening. The 700-km \( u \) at 354 K has increased slightly to 6 m s\(^{-1} \). Storm-relative \( u \), stronger than at 1200 UTC 15 September, is now evident within 500 km of the center in all quadrants but the south (Fig. 9b). As Flo has moved closer to the deepening trough in the westerlies, and the TUTT cell has migrated toward Flo's southeast quadrant, the strength of the low outflow to the north has increased relative to that in the east. This flow is also associated with a region of low PV (<0.5 PVU) at 354 K (not shown). An expanding area of outflow to the west is also evident as previously noted in the cloud pattern (Fig. 3c). At 366 K (Fig. 10b), the large-scale ridge to the west has shifted southward and a localized anticyclone has developed just north of Flo. The 700-km radius \( u \) has increased to 2 m s\(^{-1} \) with relative maxima to the southeast and southwest. The semicircle means (Fig. 11b) indicate a slight shift to higher \( \theta \) on the north side and a significant deepening of net outflow to the south.
At 1200 UTC 17 September, Flo is near peak intensity. The 354-K field (Fig. 9c) indicates outflow to the northwest and north is dominant as other sectors are being influenced by the TUTT cell to the east and the belt of westerlies to the west. As noted in section 3, the extensive cirrus associated with low outflow to the east has entirely dissipated by this time, as has the cirrus previously expanding westward over Taiwan. Winds 700 km east and west of Flo have become northerly and southerly, respectively. Though outflow to the north is stronger than at 1200 UTC 16 September, lack of significant outflow to the east and west has reduced \([u]\) to 4 m s\(^{-1}\). Strong anticyclonically curved flow along the margins of a ridge oriented east-northeast to west-southwest across the storm at 366 K (Fig. 10c) is associated with stronger outflow to the south-southwest and outflow becoming established to the north-northeast as well; \([u]\) has increased to 5 m s\(^{-1}\). These tendencies are also evident in Fig. 11c. The depth of significant \([u]\) has increased consistently with the development of outflow to the northeast noted at 366 K and above over Japan. This may possibly be a reflection of flow originating from the south side of the CDO, wrapping anticyclonically around the west side and venting north as noted in section 3. The base of \([u]\) has shifted upward to 358 K as low outflow to the east-southeast and west-southwest has largely ceased. As noted in section 3, the rawinsonde at 26°N, 126°E indicated the sudden arrival of very high (up to 382 K) outflow between 1200 UTC on 16 September and 0000 UTC 17 September.

By 1200 UTC 18 September, strong outflow develops to the south in the 354-K layer (Fig. 9d) and continues to strengthen to the north as Flo, now beginning to fill, approaches the westerlies. The region of outflow to the north also expands westward, both relative to Flo and in an absolute sense, as indicated by the cirrus edge over Korea in Fig. 3g. Low outflow \([u]\) has increased to 8 m s\(^{-1}\). High outflow (Fig. 10d) also strengthens to the south at small radii but \([u]\) weakens to 3–4 m s\(^{-1}\) as regions of inflow associated with the TUTT cell to the southeast and westerlies to the west-northwest are juxtaposed closer to the center. The semicircle means (Fig. 11d) again show consistent trends. The \([u]\) has become shallower (346–366 K), stronger at 700-km radius, and extends to larger radii. Though both “inflow” \((u < 0)\) and outflow \((u > 0)\) are associated with the westerlies (see Fig. 9d), Flo is situated in the right entrance region of a jet streak over the northern Sea of Japan and is therefore in an expected region of divergence. The \([u]\) also shows stronger outflow at lower levels, with a base at around 346 K. The decrease in \([u]\) beyond 1000-km radius is due to the strong easterlies (inflow) associated with the north side of the approaching TUTT cell (Fig. 10d).

Summarizing, the semicircle means of radial winds and maps of “low” (354 K) and “high” (366 K) storm-relative winds suggest the three-dimensional structure of Flo’s outflow evolves as follows.

- Prior to rapid deepening, outflow is manifested as a pair of “channels” and is primarily confined to levels below 358 K (about 150 hPa) with a maximum near 354 K (180 hPa). The more prominent of the two channels emanates to the north and curves anticyclonically to the east and southeast. A secondary (much weaker) channel found at slightly larger \(\theta\) emanates from the south side and curves anticyclonically to the southwest.
- As rapid deepening progresses, the strongest outflow shifts to higher \(\theta\) values, with outflow to the south beginning to occur at values of up to 378 K. Outflow at relatively low levels (354 K) continues to the north and east. A third outflow region, also predominantly low level, develops to the west of the storm.
- At peak intensity, two outflow regions are evident. The lower-level outflow to the north continues but is restricted to the east, apparently due to the circulation of the approaching TUTT cell to the east-southeast. The component of this outflow to the west has been redirected to the northwest and north as Flo begins to interact with the belt of subtropical westerlies. The second channel is a region of predominantly high level (356–382 K) outflow to the south through southwest that straddles both the static tropopause (\(\Delta P\) decreasing with height) and the dynamic tropopause (1.5 PVU), as indicated by semicircle means in these quantities (not shown) and the sounding near 20°N, 126°E. This outflow appears to contain a branch that curves sharply (anticyclonically) around the west side of the storm and is vented to the north (366–374 K).
- As Flo weakens, the two regions mentioned above become more similar in their distribution of positive radial wind with \(\theta\). The northern low outflow strengthens, and the greatest change in the south is the establishment of a strong low-level component at close radii near 350–358 K (150–250 hPa). Both north and south outflow show a similar vertical structure; by this time the high-level outflow is rotated anticyclonically about the storm relative to the low-level outflow.

6. Discussion

In section 1 it is hypothesized that the outflow from tropical cyclones should occur over an intensity-dependent range of potential temperature \(\theta\) values and should have PV lower than its environment at a given \(\theta\) value. Observations from this study show that the first hypothesis can be retained based on the results summa-
rized in Fig. 7. Evidence (Fig. 8) also appears to support the second hypothesis although the difficulty in tracing PV plumes associated with particular outflow regions questions the utility of PV as an outflow tracer, at least with the present data density.

This case represents the first description of the three-dimensional structure of tropical cyclone outflow. In addition to adding to the general knowledge of tropical cyclone structure, this description also provides a context for future research on tropical cyclone–environment interactions and on the subsequent effects on tropical cyclone structure. Results are summarized as follows.

- Outflow occurs over a 36-K range of θ during the time period analyzed. A single outflow region may contain low-θ outflow, high-θ outflow, or both. In general, the θ value at which the highest (eyewall) outflow occurs and the range of θ values over which outflow occurs increases (decreases) with increasing (decreasing) storm intensity (maximum sustained wind). During the period of storm intensification, a significant decrease in PV is also observed in the isentropic layer in which outflow is occurring.

- The thick cirrus pattern and therefore nearly all CMWs are associated with that component of the outflow with lower θ values (and therefore altitudes), in this case the 350–358 K (150–250 hPa) range. It is suggested that thin cirrus plumes associated with outflow emanating at higher levels (into the lower stratosphere) undergo rapid lateral spreading due to the high static stability (Lilly 1988); this both reduces their thickness and emissivity and increases their rate of erosion due to mixing with dry environmental (stratospheric) air, making them harder to observe and track from moderate-resolution satellite data. This issue will be addressed by the second author in future research aimed at examining full-resolution (1 km) visible satellite imagery for cirrus tracking.

- In addition to the well-known horizontal asymmetries, marked variability in the vertical distribution of outflow within the outflow layer is also frequently evident. Equatorward outflow tends to be higher, both in terms of pressure and θ, than poleward outflow in this case. Equatorward outflow is also typically confined to a smaller θ range, though not without exception.

Common practice to date has regarded the outflow layer as a vertically homogeneous slab within which the flow is well approximated by 200-hPa rawinsondes, augmented with CMWs (i.e., Merrill 1988; DeMaria et al. 1993). The results of the present study show that such a treatment is most applicable for tropical cyclones of average strength, though not necessarily for intense ones from which significant outflow may be occurring at higher altitudes (well above 200 hPa and not as evident in production-quality CMWs). It will therefore be necessary to reexamine the concepts pertaining to interactions between tropical cyclones and their environments previously based on two-dimensional analyses to include the notion of vertical structure variability. In this regard, a few preliminary observations and ideas on the three-dimensional aspects of these interactions follow.

The temporal development of outflow in different θ layers as described in section 5 is in agreement with previous studies that suggest that the orientation of horizontal asymmetries in a given θ layer is determined primarily by the large-scale storm-relative flow (Oyama 1987; Wu and Emanuel 1994). Figure 10 supports this suggestion. The anticyclonically sheared easterly flow in the 366-K layer evident before significant outflow commences (Fig. 10a) would suggest that outflow should disperse to the south and southwest, as is later observed. The direction of an existing outflow plume can also apparently be modified by changing environmental conditions. The subsequent approach (in a storm-relative sense) of the subtropical southwesterlies from the west, and the northeast flow around the TUTT cell approaching from the east, are consistent with observed outflow being mainly redirected to north and south channels, respectively, by 1200 UTC 18 September.

The area of outflow to the west that is noted at 354 K (Fig. 9) from 1200 UTC 15 September through 1200 UTC 16 September is not so easily explained by the large-scale storm-relative flow, however. Examination of the CMWs and the motion of the leading edge of the cirrus implies that this surge to the west originated on the north side of the storm prior to 0000 UTC 15 September, in an area of strong, deep easterly flow associated with a prominent convective band. It is suggested that vertical transports of momentum may have a significant influence on outflow direction, particularly in the absence of strong storm-relative environmental winds in the outflow layer.

To summarize, Flo’s outflow patterns (viewed in a planar sense) appeared to be influenced by the ambient flow during the early stages of the analysis period. Outflow was “allowed” to spread west for a time (coinciding with rapid development) into a region of weak ambient flow. Subsequently, the approach (in storm-relative terms) of the midlatitude southwesterlies and TUTT cell then “inhibited” all outflow to the west and east and “encouraged” outflow directed to the north and south. During the period after Flo’s peak intensity, an “active” interaction with strong eddy flux convergence of angular momentum at outer radii developed and the magnitude of Flo’s radial circulation at medium radii was simultaneously (on a 12-h timescale) observed to strengthen. However, Flo’s intensity (central pressure and maximum winds) continued to decrease.

In light of the findings of this study, what must be added to conceptual models of tropical cyclone–environment interactions is an account for the vertical variation of the outflow depth in relation to the environ-
mental stimuli. The vertical structure of the outflow has interesting implications on intensity change modulation by environmental influences and should be a subject of future work. In the case analyzed in this study, an intense tropical cyclone has been shown to produce significant outflow over a 36-K $\theta$ range. The hypothesis of Holland and Merrill (1984) and Molinari and Vollaro (1990) that EFC associated with a passing weather system can stimulate the secondary circulation and aid tropical cyclone development must henceforth be considered in terms of the alignment of the forcing and outflow in $\theta$ space. It is possible, for instance, that EFC concentrated at 354–362 K might assist the development of a weak tropical cyclone by enhancing its secondary circulation along its $\theta$ surface of maximum outflow (eyewall). However, as may have occurred in the later stages of Flo, this same stimulus may impede further inner-core development of an intense storm by stimulating its peripheral convection at the expense of the eyewall, from which the outflow is emanating at higher $\theta$. The effects of vertical shear in the ambient flow over the depth of the outflow layer must also be considered, as this may control whether the low-$\theta$ and high-$\theta$ outflow from an intense storm are aligned in the vertical. Because outflow is apparently associated with a negative PV anomaly, the superposition (or lack thereof) of these anomalies at different $\theta$ levels may influence the strength of the induced upper-tropospheric circulation (such features as “outflow jets”), its nonlinear evolution, and the ability of the storm to sustain outflow over its full $\theta$ range for an extended period.

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