The Structure and Evolution of a Simulated Midlatitude Cyclone over Land

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(Manuscript received 5 December 1991, in final form 13 July 1992)

ABSTRACT

Using output from a mesoscale model simulation, this paper describes the evolution of the three-dimensional temperature and humidity structures of an intense cyclone that developed over the eastern half of the United States during 14-16 December 1987. Some specific findings include the following:

- The occlusion process at the surface appeared similar to the classical paradigm, with the cold front catching up with the warm front. A loft, a warm-type occlusion structure formed as an upper baroclinic zone merged with the surface-based occluded front.
- The frontogenesis associated with the cold and warm fronts was initially continuous, but as the system evolved, a break in the frontogenesis along the northern section of the cold front developed. A single frontogenesis feature appeared to support both the warm and occluded fronts.
- A quasi-linear region of convective activity was observed in advance of the surface cold front. This convective line was associated with an upper-level humidity front, which originated from the confluence of descended, mid-, and upper-tropospheric trajectories and saturated, rising warm-sector trajectories.
- Many of the structural elements of the storm can be explained by the differing air-parcel trajectories across these features. Although most trajectories can be meaningfully grouped into a limited number of families, they cannot be presented accurately in terms of only two or three conveyor belts or airstreams.
- During the early part of the simulation, the upper and lower baroclinic zones were relatively distinct, with the upper baroclinic zone associated with an upper-level short wave-jet streak. The two baroclinic zones came together as the short wave overtook the low-level baroclinic zone.
- The model simulation of the December 1987 cyclone, as well as the observed storm itself, suggests both similarities and differences with the "T-bone" conceptual model. An attempt is made to explain these differences based on the differing environments in which the storms evolved.

1. Introduction

Although impressive gains have been made in understanding the dynamics of midlatitude cyclones and in simulating their development, a comprehensive understanding of their three-dimensional structural evolution is still lacking. Possible reasons for this situation include the following.

1) The spatial and temporal resolution of the operational radiosonde network has been insufficient for defining the detailed synoptic and mesoscale structures of cyclonic systems.

2) Attempts to improve the traditional Norwegian cyclone model (Bjerknes 1919; Bjerknes and Solberg 1922) have produced a collection of complex, and often contradictory, conceptual models. For example, anafront-katafront, trowal, split-cold-front, conveyor-belt, fractured-front, and cold-front-loft models have been proposed over the last 50 years. The generality of these conceptual models is uncertain.

3) Although the structure of upper-level fronts is now fairly well known and several mechanisms for their initiation and development have been proposed [see Keyser and Shapiro (1986) for a comprehensive review], understanding of their development within the context of midlatitude cyclone evolution is still inadequate. The relationship between upper- and lower-tropospheric fronts remains unclear.

4) Even though numerical simulation of extratropical cyclone development has improved greatly in recent years, few modeling studies have provided detailed structural analysis of simulated cyclones. Far more common is the use of models for sensitivity studies in which various physical mechanisms (such as latent heating or surface friction) are added or removed to determine their contribution to cyclone intensification.

This paper attempts to address several of the above points. First, a detailed examination of the structural evolution of a simulated cyclone over land is presented. The high temporal and spatial resolution of mesoscale model output is used to create horizontal and vertical...
cross sections, three-dimensional perspectives, and numerous air-parcel trajectories. Special attention is given to the development of the thermal and humidity structures of the storm, with intensive trajectory analysis used to understand their origins. Second, the traditional Norwegian cyclone model as well as more contemporary conceptual models are evaluated in light of the diagnosis of this storm.

This study applied the Pennsylvania State University–National Center for Atmospheric Research (PSU–NCAR) Mesoscale Model version 4 to the cyclone of 14–16 December 1987. The December 1987 storm was selected for several reasons. First, it was a vigorous, explosively deepening, midlatitude cyclone with well-developed frontal features. Second, this storm went through its complete life cycle over the eastern half of the United States, a region in which a relative abundance of observational data facilitates verification and evaluation of the model simulation. Third, this cyclone had been well forecast by the National Meteorological Center's Nested Grid Model (NGM), and thus it was expected that a higher-resolution mesoscale model would do at least equally as well. Fourth, considering that the overwhelming majority of mesoscale model investigations of midlatitude cyclone development have been for oceanic events, a land simulation allows the comparison of the structural evolutions over land and water.

In a companion paper (Schultz and Mass 1993), we describe in greater detail the occlusion process that occurred in the December 1987 cyclone and compare this occlusion with previous work.

2. Observational overview of the 14–16 December 1987 storm

This section provides a terse observational description of the development of this storm, both to acquaint the reader with the cyclone's general features and to provide a basis for verifying the model results presented in section 4. Further information about this event, specifically regarding the occurrence of large-amplitude gravity waves, can be found in Schneider (1990).

At 1200 UTC 14 December 1987, a weak surface low center was observed over southern Texas along a stationary front, which was the extension of a cold front that had passed through the eastern United States the previous week (not shown). At 500 mb, a short-wave trough over northwest Mexico was embedded within a longer wave trough covering much of the western United States. At this time, the largest baroclinicity at 500 mb was associated with the short wave and was quite distinct from the low-level baroclinicity over Texas and the southeast United States.

By 0000 UTC 15 December, the storm had deepened by a few millibars at the surface and the surface low center appeared to be displaced slightly back into the cold air (Fig. 1a). At 850 mb (Fig. 1b), the baroclinicity associated with the warm and cold fronts was quite apparent, and at 500 mb (Fig. 1c), the short-wave trough (and accompanying large upper-level temperature gradient) had sharpened and progressed into western Texas.

During the next 12 h, the storm deepened explosively, with the central pressure dropping by approximately 23 mb. The surface analysis at 1200 UTC 15 December (Fig. 2a) suggests that the storm had begun to occlude. With a section of the surface warm front being retarded by cold-air damming on the eastern side of the Appalachians, a region of warm air was becoming isolated (secluded) over southern Ohio and eastern Kentucky. In contrast, at 850 mb (Fig. 2b), there is less suggestion of topographic effects. At 500 mb (Fig. 2c), the short-wave trough had closed off and was located almost directly above the surface low pressure center. Dry air aloft was observed above much of the warm sector over the southeast United States, and the upper-level baroclinic zone at 500 mb was approaching the frontal zone in the lower troposphere.

By 0000 UTC 16 December, the surface low center had begun to fill (to approximately 985 mb), cold-air damming to the east of the Appalachians had greatly attenuated, and a secondary low center had formed near the intersection of the warm, cold, and occluded fronts (Fig. 3a). At 500 mb, the upper-level front was considerably attenuated (Fig. 3c).

Geostationary satellite imagery for a portion of this event is presented in Fig. 4. At 0001 UTC 15 December, a comma-shaped cloud mass is evident, the tail of which contained active convection. By 1101 UTC 15 December, the area of high clouds had expanded and the line of convection extended from West Virginia southwestward to the Gulf of Mexico. Finally, at 1701 UTC, when the system was occluded, the higher clouds appeared to be split between a band circling into the low center and a convex cloud shield to the southeast, with a slot of nearly cloud-free skies extending towards the low center. Note that the clouds were positioned in advance of the cold and occluded fronts, thus underlining the dangers in using the back edge of upper cloud features for determining frontal positions. Later in this paper, numerical model diagnostics will be used to examine the origin of several of these satellite features.

1 It is interesting to note that over the eastern half of the United States, the most “classic” cyclone developments occur under such flow patterns, that is, with a long-wave trough over the western United States and accompanying southwesterly flow over the central part of the country. Such a large-scale flow environment minimizes the topographic effects of the Rocky Mountains, which can be quite significant for more westerly flow regimes.

2 An occlusion is defined as back-to-back cold- and warm-frontal zones without an intervening warm sector. It is associated with a narrow tongue of warm air, a surface pressure trough, and a distinct veering of the surface wind.
Composite radar imagery for this storm is shown in Fig. 5. At 0035 UTC 15 December, a north–south convective band was closely aligned with the surface cold front, with some tops reaching 40 000 ft (12 192 m). By 1135 UTC 15 December, the convective band had generally weakened, although a section near the Gulf coast remained vigorous. A large area of stratiform precipitation was apparent north and east of the surface occluded and warm fronts. Finally, at 1735 UTC, the convective band was still evident and had moved to
the Atlantic coast, while the large region of stratiform precipitation associated with the low center and occluded front had begun to dissipate.

In summary, during this event a strong upper-level short-wave trough moving through a long-wave trough interacted with a preexisting baroclinic zone (and coincident weak surface trough) to produce an explosively deepening cyclone. Additional features included
an upper-level frontal zone, cold-air damming, secondary low development, mesoscale gravity waves, and a long-lived convective line.

3. Description of the PSU–NCAR Mesoscale Model

To produce a model "dataset" of the evolution of the 14–16 December 1987 event, the PSU–NCAR Mesoscale Model version 4 (i.e., the MM4) was used. This three-dimensional, primitive equation, hydrostatic, mesoscale model is described in Anthes et al. (1987) and Zhang et al. (1988). The MM4 uses a terrain-following sigma (σ) vertical coordinate system.\(^3\)

\(^3\) Here \(\sigma = (p - p_s)(p_s - p_t)^{-1}\), where \(p\) is pressure, \(p_s\) is surface pressure, and \(p_t\) is the constant pressure of the top of the model (100 mb).
Fig. 4. Enhanced infrared geostationary satellite imagery at 0001, 1101, and 1701 UTC 15 December 1987. Surface fronts are also shown.

with the horizontal boundary conditions determined from interpolation of 12-h National Meteorological Center (NMC) operational analyses. The parameter-
izations of the surface and planetary boundary layers in the MM4 were developed originally by Blackadar (1979) and are described in Zhang and Anthes (1982). Nonconvective precipitation is calculated using explicit prognostic equations for water vapor, cloud water, and rainwater (Hsie et al. 1984). Convective precipitation was parameterized using an Arakawa–Schubert (1974) scheme, as modified by Grell et al. (1991) to include the effects of convective-scale downdrafts. Testing four convective precipitation parameterizations on the MM4, Kuo and Low-Nam (1990) concluded that the above Arakawa–Schubert scheme produced the most realistic simulations. The MM4 and its predecessors have proven themselves capable of realistically simulating cyclone development for a variety of regions and synoptic conditions (Anthes 1990).

Sixteen vertical levels were used in this simulation.4 Within the planetary boundary layer, the sigma-layer thicknesses averaged approximately 20 mb, while in the upper domain they were 90 mb in vertical extent. The horizontal grid spacing was 45 km on an array of 107 × 91 grid points. The model domain covered the continental United States, southern Canada, and northern Mexico. The model was initialized at 1200 UTC 14 December 1987 and ran for 36 h until 0000 UTC 16 December 1987 using a 70-s time step, with output being stored at full resolution every 15 min. The model analysis–initialization began with the NMC final analysis. This analysis was then modified to fit surface and upper-air observations using a Cressman scheme. A limited amount of bogusing was used to more accurately depict the low-level temperature gradients. The model initialization scheme assumed zero net mass divergence in the column.

4. Simulation description and verification

a. Cyclone intensity and position

Figure 6 presents the simulated and observed (analyzed) central (sea level) pressures of the cyclone for the 36-h simulation (from 1200 UTC 14 December to 0000 UTC 16 December). During the first 12 h, a period in which only weak intensification took place, the model central pressure fluctuated around the analyzed NMC values with errors of several millibars. From 12 through 21 h, the modeled and observed central pressures were quite consistent, with both indicating explosive development (approximately 15-mb drop during the 9-h period). Modeled and observed central pressures diverged during the next 9 h (21–30 h) as the observed cyclone filled, while the model storm continued to deepen. During the final 6 h, both observed and modeled cyclones weakened.

b. Horizontal structure

Figure 8 presents model heights (or pressures) and temperatures at the surface, 850, and 500 mb at 12, 24, and 36 h into the simulation (corresponding to 0000 UTC 15 December, 1200 UTC 15 December, and 0000 UTC 16 December, respectively). At 12 h into the simulation the height, pressure, and temperature fields are very similar to the observed5 (cf. Figs. 1a–c). The simulated system is an open wave at the surface, with both the observed cold-air damming and upper-level frontal zones well represented.

By 24 h, the surface warm sector had narrowed substantially and the model cyclone had just begun to occlude. The model output contrasts with the surface analysis for this time (Fig. 2a), in which a substantial occluded front was apparent. This discrepancy resulted from the slower forward motion of the simulated cold front compared with the observed. At 850, 700 (not shown), and 500 mb, the model and observed low centers were in close proximity, and the only significant difference in intensity was at 850 mb where the sim-

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4 That is, σ = 0.0, 0.1, 0.2, 0.3, 0.4, 0.5, 0.6, 0.7, 0.78, 0.84, 0.89, 0.93, 0.96, 0.98, 0.99, and 1.0.

5 In considering the comparisons between the NMC upper-air analyses and the model output, one should note that the former often underestimate maxima and minima and the intensity of horizontal gradients.
shown are surface frontal positions based on model output. At 850 mb, a comma-shaped cloud mass is apparent throughout the period. As the simulation proceeds, a tongue of dry air projects northeastward behind the cold front. At 500 mb, a large area of dry air sweeps eastward during the simulation. The leading edges of the dry air (and the back edge of the model clouds) are positioned far east of the cold front at 24 and 36 h and in advance of the occluded front at 36 h, again illustrating the dangers inherent in positioning fronts using the back edge of upper-level cloud features.

In order to better illustrate the three-dimensional moisture structure of the storm, a three-dimensional perspective drawing of the saturated (100% relative humidity) volume of the cyclone has been created at 30 h, when the system had completed its explosive development and was occluded (Fig. 10). This view from the south reveals the dry slot extending toward the storm center at low levels and the forward-leaning topology of the saturated volume to the north and east of the low center. The surface cold front is positioned near the rear of the shallow cloud deck, with deep moisture hundreds of kilometers to the east associated with the convective line. As discussed in section 6, this moisture distribution is the result of dry air aloft streaming over the system and is reminiscent of the split-cold-front model first discussed by Browning and Monk (1982).

A useful, and all too rarely applied, measure of model verification is precipitation amount. Figure 11 presents the total precipitation produced in the model storm for 27–30 h into the simulation (nominally 1500–1800 UTC 15 December 1987). Total precipitation evinces a mushroom-shaped pattern, with the heaviest precipitation over the Great Lakes and from coastal South Carolina to the central Florida coast. The latter feature is mainly of convective origin. Compared to the radar reflectivity at 1735 UTC (Fig. 5), the model appears to be producing too much precipitation in the northern “head” and the convective line is displaced approximately 150 km too far to the east. Surface precipitation observations (not shown) confirm that although the model precipitation pattern is generally realistic, the amounts, especially in the comma head, are generally 50%–100% too large.

5. Structural analysis of the simulated cyclone
a. Surface temperature evolution

Surface temperature analyses using a 1°C contour interval were created to delineate the storm’s low-level thermal evolution (Fig. 12). At 12 and 18 h into the simulation, the thermal patterns indicate open waves at the surface, with relatively well-defined cold and warm fronts. The most intense temperature gradients were located within the frontal zone in the vicinity of the low center. By 24 h, the warm sector had narrowed and occlusion had just begun as the cold front rotated
Fig. 8. Temperature (°C) and sea level pressure (mb) at the surface and temperature (°C) and geopotential heights (m) at 850 and 500 mb for 12, 24, and 36 h into the simulation. Contour intervals are 4°C, 4 mb, 30 m, and 60 m for temperature, sea level pressure, and 850- and 500-mb geopotential heights, respectively.
FIG. 9. Relative humidity (%) at 850 and 500 mb at 12, 24, and 36 h into the simulation. Surface frontal positions, based on model output, are also displayed. The contour interval is 20%, and saturated regions (RH > 100%) are shaded.
counterclockwise around the low, and the western section of the warm front slowly shifted northward. The eastern section of the warm front, retarded by the Appalachians, remained nearly stationary. The influence of the Gulf of Mexico in weakening the temperature gradient behind the cold front is also evident. At 27 h, the occlusion process at the surface is clearly evident near the low center, with the cold front making contact with the slower-moving warm front. During the remainder of the simulation, the cold front continued to advance toward the stalled warm front on the west side of the Appalachians, removing warm air from the surface. As discussed in section 6, the thermal evolution displayed in Fig. 12 differs substantially from the fractured-front–T-bone structures described in Shapiro and Keyser (1990). For example, there is little evidence of fracturing of the cold front, back-bent warm fronts, or the seclusion of relatively warm air in the vicinity of the low center.

b. Vertical cross section through the occluded cyclone

In order to document the three-dimensional structure of the simulated cyclone after occlusion, vertical cross sections of temperature and relative humidity (Fig. 14) were constructed through the storm at 30 h into the simulation at various distances from the low center (locations of the cross sections are shown in Fig. 13). Lines have been placed on the temperature cross sections to highlight discontinuities in temperature gradient; these lines should not be assumed to be the same as classical Norwegian frontal symbols.

The southernmost cross section (A) has its vertex well within the warm sector. The corresponding temperature field shows a wide warm sector of relatively uniform temperature, a warm front sloping to the right, and a complex cold-frontal structure in which an upper- to midtropospheric baroclinic zone appears to mesh with a surface-based frontal zone. One notes that the forward edge of the combined frontal zone tilts forward to approximately 750 mb. This forward-tilting feature is associated with a core of large vertical motions, centered near the point of maximum forward extension. Large adiabatic cooling or diabatic effects (such as evaporation) are possible mechanisms for establishing this nonclassical feature. Aloft, dry air extends east of the surface cold front and descends to the surface in a narrow tongue behind the cold front.
Fig. 12. Surface temperature evolution (1°C contour interval) at 12, 18, 24, 27, 30, and 33 h into the simulation (nominally 0000 UTC 15 December through 2100 UTC 15 December 1987). Model surface low positions and the locations of the vertical cross sections displayed in Schultz and Mass (1993) are also shown.
Cross section B cuts the system along the southern part of the occluded front. A warm sector (with relatively weak horizontal temperature gradients) is not apparent below 850 mb. An upper-tropospheric baroclinic zone extends into the lower troposphere and appears to contribute what would traditionally be classified as the upper cold front of a warm-type occlusion. A surface-based cold-frontal zone is beneath the upper baroclinic zone. As in section A, dry air extends into and above the warm sector; a tongue of this dry air extends toward the surface behind the occluded front.

The northernmost cross section (C) cuts through the occlusion not far from the low center. The thermal pattern in this cross section also resembles a warm-type occlusion. The occlusion’s warm tongue and flanking horizontal temperature gradients are substantially attenuated (compared to section B), and the warm sector aloft has shrunk considerably. The occluded frontal complex is embedded within the saturated air, except for a narrow tongue of dry air trailing immediately behind the upper-level frontal zone.

c. Frontogenesis

During the past few decades it has become apparent that in contrast to the classical Norwegian model, midlatitude cyclones often develop within relatively diffuse baroclinic zones that sharpen into frontlike structures during cyclogenesis. This section documents surface frontogenesis during the evolution of the model cyclone using the Lagrangian frontogenesis function of Miller (1948). This function calculates the total time derivative of the absolute magnitude of the horizontal potential temperature gradient,

\[ F = \frac{d}{dt} |\nabla \theta|, \]

using the model wind and temperature fields. The diabatic contribution to frontogenesis was neglected in calculating the fields presented below.

Figure 15 presents the surface frontogenesis fields and model surface frontal positions at four times during the simulation. At 12 h, frontogenesis was found within the cold- and warm-frontal zones, with the largest values located along the warm front immediately to the north and east of the low center. Six hours later (18 h), frontogenesis had intensified, being especially strong near the surface low center. This region was characterized by both strong temperature gradient and confluence as the peak of the warm sector extended northward to meet the eastward-flowing cold air north of the low. A subtle, but significant, feature at this time is the weakening of frontogenesis along the northern section of the cold front. At 24 h, the “fracture” between the strong frontogenesis along the warm front and the secondary maximum along the cold front had become more pronounced. Three hours later (27 h), at which time the system was occluded at low levels, the break between the frontogenesis along the middle portion of the cold front and the frontogenesis associated with the warm–occluded front was nearly complete. A single frontogenesis feature supported both the warm and occluded fronts, with the occluded front being a zone of active (if not intense) frontogenesis. The significance of both this fact and the fracturing of frontogenesis along the cold front will be explored in section 6d.

d. Three-dimensional trajectories

Trajectory analysis provides a powerful tool for understanding the origins of the structures observed in midlatitude cyclones. To that end, thousands of forward and backward trajectories were constructed for the 14–16 December 1987 cyclone using 15-min output from the numerical simulation and applying the trajectory routine described in Seaman (1987). In this section, backward trajectories ending along a vertical cross section and on a series of quasi-horizontal cuts will be presented.

1) Trajectories ending on a vertical cross section through the storm at the initiation of occlusion

This section explores the origins of the temperature and humidity structures appearing in a vertical cross section through the storm at 24 h (immediately after
FIG. 14. Vertical cross sections of temperature (°C) and relative humidity (%) at 30 h into the simulation. The temperature contour interval is 2°C and relative humidities greater than 90% and less than 40% are indicated by light and dark shading, respectively. The locations of the cross sections are shown in Fig. 13.
Fig. 15. Surface frontogenesis of the model storm at 12, 18, 24, and 27 h into the simulation. The contour interval is $0.2 \times 10^{-4} \, \text{K m}^{-1} \, \text{s}^{-1}$ [or approximately $2^\circ \text{C (100 km)}^{-1} (3 \text{ h})^{-1}$]. Surface frontal positions, based on model output, are also shown.

Model surface frontal positions at 24 h are shown (solid lines). Because the 3-h baroclinic zone was relatively diffuse, the 800-mb $10^\circ \text{C}$ isotherm (which lies on the warm side of the baroclinic zone) at 3 h is shown for reference (dashed lines). When examining the trajectories, a comparison with the 3- and 24-h temperature fields presented in Fig. 23 is often useful.

At the lowest model level ($\sigma = 0.995, \sim 950-970$ mb), where the air parcels experienced relatively little vertical motion, the baroclinicity associated with the cold front originated in the confluence of air parcels of differing trajectories and initial temperatures. Trajectories 1–5 (trajectory 4 not shown) originated deep within the cold air and approached the low center as it tracked to the northeast. In contrast, trajectory 7,

initiation of occlusion) by examining backwards trajectories from 24 to 3 h into the simulation. The cross section, the location of which is shown in Fig. 12, presents the ending positions of the backwards trajectories and relative humidity, as well as lines highlighting breaks in horizontal temperature gradients (Fig. 16a).

The trajectories ending within the 24-h cross section are presented in Figs. 16b–e. In order to provide information regarding the height of the trajectories, each is displayed using a convention in which the width of the trajectory varies according to its pressure level.

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6 More detailed trajectory analyses during the occluded stage are found in the accompanying paper (Schultz and Mass 1993).
FIG. 16. (a) Vertical cross section of temperature and relative humidity at 24 h into the simulation. The temperature contour interval is 2°C and relative humidities greater than 90% and less than 40% are indicated by light and dark shading, respectively. The location of the cross section is shown in Fig. 12. Also shown are the ending positions of a number of backward trajectories (34 to 3 h). Backward trajectories from 24 to 3 h that ended in the cross section at the $\sigma = 0.995$ (b), 0.795 (c), 0.595 (d), and 0.395 (e) model levels. The 24-h model surface frontal positions (solid) and the 800-mb 10°C isotherm at 3 h (dashed), which is located on the warm side of the baroclinic zone at that level, are also shown. The width of the trajectories is proportional to pressure (see insets) and an arrowhead is shown within each trajectory at 3-h intervals.
ending on the warm side of the cold-frontal zone, began far to the south within the warm air and passed through the front because of frictional retardation near the surface. Trajectories that terminated in the warm sector (trajectories 8–11, 9 not shown) originated within the baroclinic zone along the Carolina coast. The potential temperatures of these trajectories increased substantially ($\sim 14^\circ$C) as they crossed the Appalachians into the warm sector as a result of mixing. Trajectory 12, terminating within the warm-frontal zone, originated on the northern side of the low-level baroclinic zone, while trajectory 13 remained deep within the cold air throughout its entire journey.

Terminating at the $\sigma = 0.795$ level ($\sim 800–775$ mb), trajectories 27 and 28 began and ended deep in the cold air and evinced little vertical motion. Trajectories 29–31 (30 not shown) originated within the initial baroclinic zone and rose as they were swept around the low into the cold-frontal zone. Trajectory 32 began (at 919 mb) within the baroclinic zone over Texas and ascended during the period. It serves as a transition to trajectories 33 and 34, which originated far to the southwest and descended into the baroclinic zone. These two descending trajectories explain the tongue of low relative humidity at this level. Another major trajectory break occurs with trajectories 35–37 (37 not shown), which began deep in the warm air, rose vigorously (over 100 mb), and terminated within the warm sector. Trajectory 38 began in the warm sector aloft and moved northward with relatively minor (approximately 70 mb) downward displacement. In contrast, trajectory 39 subsided by 125 mb as it maintained an east-northeasterly heading in the cold air north of the warm front.

At the $\sigma = 0.595$ level ($\sim 625–605$ mb), trajectories 53–55 (54 not shown), which began in the lower-tropospheric baroclinic zone and ended within the cold-frontal zone, rose rapidly (by 120–300 mb) as they were swept counterclockwise around the low center. Trajectory 56 began in the warm air and was drawn toward the low center where it experienced intense lifting (over 350 mb). It serves as a transition to a family of dry, subsiding trajectories (57–59, 57 not shown). As will be discussed in section 6, the differential subsidence of these air parcels contributed to the existence of the upper-level baroclinic zone. Proceeding into the warm sector, one finds a very different group of trajectories (60–64, 62–63 not shown) that rose rapidly from origins deep in the warm air. Moving farther northeast in the cross section, one finds a trajectory (65) that, like trajectory 39 below, drifted to the northeast.

The final line of backward trajectories terminated at $\sigma = 0.395$ ($\sim 445–435$ mb). On the cold side of the upper-level front, trajectories 79–81 (79 not shown) rose rapidly as they approached and were swung around the low center; the first two began in the midtroposphere (707 and 629 mb, respectively), while the latter (81) started near the surface (953 mb) within the baroclinic zone. In contrast, trajectories 82–84 all subsided cyclonically into the upper-level frontal zone (from 381, 285, and 229 mb, respectively) from the upper troposphere and lower stratosphere. The trajectories that ended in the dry zone protruding above the warm sector (85–89, 86–88 not shown) all experienced net subsidence. Ending in the warm, moist air of the eastern warm sector, trajectories 90 and 91 rose from the midtroposphere (718 and 646 mb, respectively).

It appears that many of the temperature and relative humidity structures in the vertical cross section at 24 h can be explained by the differing air-parcel trajectories across these features. An attempt to summarize these findings is presented in Fig. 17, which characterizes the nature of the trajectories ending at various locations along the cross section. Thin solid lines are used to separate general trajectory families.

**FIG. 17.** Schematic summary of the trajectories ending within the vertical cross section at 24 h presented in Fig. 16a. Temperature (°C) within the vertical cross section is also shown.
Behind the surface-based cold front, the air experienced little vertical motion as it circled around the low center (A). The cold-frontal zone air (B) originated mainly within the warm frontal zone and rose as it was swept around the low center. An upper-level frontobaroclinic zone (C) protruded downward to meet the forward section of the surface cold front. This upper front was associated with descended trajectories that began in the middle to upper troposphere and lower stratosphere, with the warm side of the upper front associated with the greatest subsidence. Between the upper front and the surface-based cold front, a transitional region (D) was associated with a variety of trajectories—some began in the warm-frontal zone and rose rapidly as they circled around the low center, while others began far to the west and approached the system with little vertical motion. The nominal warm sector of the system was hardly uniform in the nature of the trajectories that terminated there. A zone of low relative humidity (E) extended eastward above the surface warm sector and was associated with previously descended trajectories from the west and southwest. Many of these trajectories were rising at the ending time (24 h) but were too dry from previous descent to produce saturation (Danielsen 1966a,b; Durran and Weber 1988; Kuo et al. 1992). Much of the lower and eastern part of the warm sector (F) was composed of moist, rising trajectories (F) that originated at low and middle levels in the warm air. Trajectories that terminated in the warm-frontal zone (G) generally remained within the zone throughout the entire simulation. A final group of trajectories (H) was composed of cold air parcels that began and remained within the cold air north of the front; these air parcels began in the preceding anticyclone and experienced little vertical displacement. Although the trajectories presented in this section can be meaningfully grouped into a limited number of families, it is equally clear that they cannot be accurately presented in terms of only two or three conveyor belts or airstreams. This point will be amplified in the discussion section of this paper.

2) Trajectories Ending on Quasi-Horizontal Planes after the Initiation of Occlusion

In order to describe more comprehensively the three-dimensional airflow of the December 1987 cyclone, backwards trajectories from 27 h (at which time the model cyclone had begun to occlude) to 12 h that terminated on four different pressure levels are presented in Fig. 18. Initial (dotted) and final (dashed) surface frontal positions based on model output are also shown in this figure.

At 900 mb, the cold front (at 27 h) separates rising warm-sector trajectories from the southeast (5, 12, and 19) and dry descending trajectories from the southwest (4 and 25). It is necessary to go several hundred kilometers behind the cold front to find trajectories that began deep within the cold air (10 and 24). These cyclonically turning trajectories rose as they traveled around the low. There is a confluence of trajectories of widely varying origins across the surface occluded front. At and north of the front, the air parcels (31) move northward and originate within the initial warm-frontal zone. To the south of the occluded front there are rising cold (24) or subsided (but now rising) dry (25) trajectories. [Schultz and Mass (1993) describe the trajectories across the occluded front in greater depth.] Finally, deep within the cold air to the north there is a class of trajectories characterized by slow eastward drift (e.g., 42 and 56).

Terminating at 700 mb, a group of ascending southerly trajectories (19, 33, and 39) began within the warm sector at 12 h, rose rapidly, and finished above the warm sector and north of the warm front at 27 h. Air parcels that started deep within the cold air to the north of the initial warm front (48 and 56) drifted eastward while subsiding gently. Other air parcels starting somewhat lower or farther to the west rotated cyclonically around the low center (8 and 22). Finally, there is a group of northeastward-moving trajectories that descended rapidly from the middle (4) and upper (25) troposphere to the vicinity of the surface cold front.

At 500 mb, dry, descended trajectories starting out well to the west became more dominant. The northernmost trajectories of this group (e.g., 25) subsided by over 200 mb during the first 12-h period and subsequently rose gently. In contrast, trajectories farther to the south (4) subsided only one-third as much. Some trajectories (22) began near the surface within the warm sector and rose rapidly as they were caught in the cyclonic circulation of the storm. Other trajectories that began above the warm sector at higher levels (42) were swept anticyclonically to the northeast. Finally, there are several trajectories (50 and 56) that drifted to the northeast with little vertical motion.

The trajectories ending at 300 mb can be divided into two major groups: 1) a collection of unsaturated trajectories with origins far to the west (e.g., 4, 7, and 28) that terminated over the southern part of the domain and 2) a group of saturated, rising trajectories with origins within the warm air (e.g., 22, 36, and 56). One notes how these warm trajectories fan out in a way that does not resemble a conveyor belt.

6. Discussion

In this paper we have provided a description of the evolving structures and airflow of the 14–16 December 1987 cyclone. This section shall further analyze some of the significant features of this storm and compare the model simulation to well-known conceptual models of cyclone evolution. A more detailed discussion of the occlusion process is presented in the accompanying paper (Schultz and Mass 1993).
a. The upper-level humidity front and its relationship to prefrontal convective lines

In this cyclone, as in many others, a quasi-linear region of convective activity (i.e., a prefrontal squall line) was observed in advance of the surface cold front. During the 14–16 December 1987 event, the convective line and the cold front were originally collocated, but as the storm evolved, a separation between the two features developed. An enhanced infrared satellite image at 2301 UTC 15 December 1987, on which the NMC surface frontal positions are superposed (Fig. 19), shows that at this time the convection was located far to the east of the surface front.

The model simulation captured both the convective line and its phasing relative to the surface cold front. Figure 20 presents the total precipitation produced by the model from 33 to 36 h into the simulation, as well as surface frontal positions (based on model output) at 33 h. The model, as in the real world, positioned the heaviest precipitation well in advance of the surface cold front.

A northwest–southeast vertical cross section of potential temperature and relative humidity through the cold front and the prefrontal convective line at 33 h into the simulation is shown in Fig. 21 (see Fig. 20 for the cross-sectional position). The potential-temperature cross section suggests that the surface-based cold front joined with an upper-level front–baroclinic zone that extended from the tropopause. The corresponding relative humidity distribution shows a large region of dry air in the upper and midtroposphere that reached the surface immediately behind the surface cold front. Aloft, this dry pool protruded far to the east of the surface cold front into a region of weak horizontal temperature gradient. We term the boundary between
the mid- and upper-tropospheric dry air and moist warm air the upper-level humidity front [McBean and Stewart (1991)] termed this feature the upper-level moisture front]. The position of the forward edge of the simulated humidity front in the mid–upper troposphere is nearly coincident with the aforementioned convective line. Finally, the region of lower humidity on the eastern edge of the cross section originated in the dry descended air behind a cold-frontal zone that had passed through the region several days earlier.

Insight into the origin of the humidity front shown in Fig. 21 can be gleaned from a line of backwards trajectories (from 33 to 12 h into the simulation) that end at the $\sigma = 0.595$ level (~600 mb) within the cross section. These trajectories are presented in Fig. 22 and their ending points are indicated in Fig. 21. Trajectory 45, which remained in the midtroposphere for its entire lifetime, was relatively dry (45% RH) at its endpoint. In contrast, trajectory 44, which originated at low levels deep within the warm sector, rose rapidly during this period (from 956 to 643 mb) and ended within the tongue of saturated air. Just within the humidity front, trajectory 43 ascended far less than 44 and finished with a relative humidity of approximately 50%. In sharp contrast, trajectory 42, which finished west of the upper humidity front but east of the surface cold front, had an origin far to the west and descended from 540 to 636 mb during its journey. This trajectory terminated with a relative humidity of only 4%. Trajectories 40 and 41, which ended in the upper frontal zone, descended rapidly from far to the west (trajectory 40 subsided from 309 to 620 mb during the period!) and terminated with extremely low relative humidities. Based on the above trajectories, it appears that the upper humidity front was established by the confluence of the descended trajectories from the west with saturated, rising warm-sector trajectories originating deep in the warm air.

The upper-level humidity front described above is reminiscent of the split-cold fronts described by Browning and Monk (1982), Young et al. (1987), and Browning (1990). As noted in these references, the upper-level part of the split front is the leading edge of dry, low-$\theta_e$ ($\theta_e$) air advancing aloft ahead of the surface cold front. Usually the upper feature is only apparent in the humidity field, being the boundary between the rising moist air of the warm conveyor belt and the descended dry air to the west. Underlying the dry air aloft is a shallow, moist zone. The upper-level humidity front diagnosed in this model simulation possesses all of these split-cold-front characteristics.

A reasonable hypothesis regarding the origin of the convective line is that the superposition of dry, lower-$\theta_e$ air aloft over moist, higher-$\theta_e$ flow below produced
a convectively unstable environment. This instability was released by the synoptic-scale upward motion that existed in that region. A cross section of equivalent potential temperature (not shown) indicates that equivalent potential temperature decreased with height near the leading edge of the humidity front aloft.

Fig. 21. Vertical cross section of potential temperature (K) and relative humidity (%) at 33 h along the path indicated in Fig. 20. The temperature interval is 2 K and relative humidities greater than 90% and less than 40% are indicated by light and dark shading, respectively. The ending points of a series of backward trajectories are also shown.

Fig. 22. Backward trajectories (from 33 to 12 h into the simulation) using model data that terminated along a line at \( \sigma = 0.595 \). The location of the line is shown in Figs. 20 and 21. Surface model frontal positions at 12 (dashed) and 33 h (solid) are also shown.

The release of convective instability produced by the superposition of a descended dry airstream over moist, warm air below does not necessarily have to occur at the leading edge of the dry airstream aloft. For example, Carr and Millard (1985) discussed the initiation of a convective line along a surface dry line within the postcold frontal dry slot of an occluded system.

b. Major airflows through the cyclone and their relationship with conveyor-belt models

Several papers (e.g., Eliassen and Kleinschmidt 1957; Browning and Harrold 1969; Harrold 1973; Carlson 1980; Golding 1984; Browning 1986; Young et al. 1987; Kurz 1988; Browning 1990) have described the major airflows through midlatitude cyclones in terms of a limited number of discrete airstreams or “conveyor belts.” Using relative-flow isentropic analysis as their basic tool, these studies have generally noted three airstreams. First, there is a warm conveyor belt (WCB) associated with most of the cloudiness and precipitation of the cyclone that begins at low levels within the southern portion of the warm sector and climbs anticyclonically above the warm front. Second, a cold conveyor belt (CCB) originates within the anticyclonic low-level flow to the northeast of the cyclone and extends westward (relative to the eastward-moving cyclone) north of the surface warm front, undercutting the warm conveyor belt. According to Carlson (1980), the cold conveyor belt rises and emerges from beneath the western edge of the warm conveyor belt (producing the western extension of the comma head) and then ascends anticyclonically to merge with the WCB.

7 Except Golding (1984), who analyzed numerical model output.
Browning (1990) suggested that part of the CCB can also descend cyclonically around the cyclone center to a position behind the cold front. The final major air flow, the dry airstream or dry intrusion, descends cyclonically from the upper troposphere or lower stratosphere into the lower troposphere above and behind the developing cyclone. Browning (1990) suggested that surface cold fronts and upper cold (humidity) fronts delineate boundaries between the dry airstream and the warm conveyor belt.

Diagnosis of the December 1987 storm simulation provides a challenging test of the conveyor-belt approach. For example, while most conveyor-belt studies have used only a handful of isentropic surfaces to examine the airflows through cyclones, this study considered several thousand trajectories (both backward and forward) launched from numerous levels and at several times. The examination of model trajectories also does not assume (as does the relative-flow isentropic approach used in the conveyor-belt papers) that the system in question translates without change of shape or speed. Finally, the trajectories based on the model simulation possess high temporal and spatial resolution (15 min, 45 km), in contrast to the coarse network of twice-daily radiosonde used in most conveyor-belt investigations.

A review of the model trajectories presented in Figs. 16 and 18 [as well as others shown in Schultz and Mass (1993)] suggest many major deficiencies with the conveyor-belt model, at least in the context of the 14–16 December 1987 event. Some examples include the following:

1) The air motions through the cyclone cannot be neatly divided into three airflows (i.e., the WCB, the CCB, and the dry airstream). Some illustrations:

- The air within and behind the cold front at 24 h (Fig. 16) had a multiplicity of origins and was not simply descended dry air as suggested by Carlson (1980) and Browning (1990). Specifically, some of the air began north of the warm front and ascended cyclonically into the cold-frontal zone (trajectories 29, 31 in Fig. 16), while other trajectories (33 and 34) descended into the frontal zone. Far behind the cold front, the air evinced little vertical motion at all (trajectories 1–3). A few trajectories that ended within the cold-frontal zone near the surface even began in the warm sector, being overtaken by the advancing surface front.

- Some trajectories (e.g., 56 of Fig. 16) began in the warm sector at low levels and were rapidly thrust aloft into positions behind the upper-level baroclinic zone. A considerable number of trajectories began in the midtroposphere north of the warm front and simply drifted eastward with relatively little vertical motion (trajectory 39 of Fig. 16 or trajectories 48 and 56 at 700 mb in Fig. 18).

- Not all dry (i.e., low RH) trajectories possess similar histories, and thus they cannot be considered one airstream. While it is true that the driest trajectories began in the upper troposphere or lower stratosphere and descended cyclonically into the mid- and lower troposphere (e.g., 84 of Fig. 16), other parcels (89) began far to the southwest and approached the system with substantially less vertical displacement. These important trajectory variations are not considered in the conveyor-belt model.

2) The model trajectories for the December 1987 storm do not suggest an anticyclonically ascending cold conveyor belt as presented by Carlson (1980). In this storm, most of the air parcels beginning north of the warm front in the lower troposphere turned cyclonically and ended within the cold-frontal zone. In Carlson (1980), the head of the comma-cloud pattern was formed by the rising cold conveyor belt; in contrast, for the December 1987 storm, the comma head was produced by rising, cyclonically turning trajectories with origins at low levels in the warm air (cf. Fig. 18 at 300 mb).

3) The warm-sector trajectories (corresponding roughly to the warm conveyor belt) do not solely possess anticyclonic turning as they rise into the upper troposphere. In fact, the rapidly rising trajectories based in the warm sector appear to fan out with some turning cyclonically toward the west and others curving anticyclonically toward the east. Kurz (1988) found a similar fanning out of the warm-sector trajectories.

4) The airflow families that do exist are not beltlike in structure, but rather evince more varied and complex geometries.

It appears that the conveyor-belt approach provides a coarse description of the nature of the air-parcel trajectories associated with midlatitude cyclones, with high-resolution (both in time and space) model trajectories suggesting that the true nature of the flows is substantially more complex. Even with its substantial drawbacks, the conveyor-belt approach has had a beneficial influence by encouraging a shift in the synopticians' viewpoint away from preexisting frontal zones as the seat of most weather toward a greater consideration of the evolutions and interactions of the major airflows of midlatitude cyclones.

c. Upper-level frontal development within the context of the simulated cyclone evolution

Although there is an extensive literature on the structure and development of upper-level frontal features [see Keyser and Shapiro (1986) for a comprehensive review], only a handful of studies (e.g., Reed 1955; Bosart 1970; Mudrick 1974; Hines and Mechosso 1991; Sanders et al. 1991) have examined the development of such frontal zones in the context of the three-dimensional evolution of baroclinic systems. As a re-
sult, several major issues regarding upper-level frontal development are still outstanding, such as the relative importance of horizontal confluence versus differential vertical motions in producing these features and the relationship between upper-tropospheric and surface-based frontal zones. This simulation provides an excellent opportunity for applying the high temporal and spatial resolution of a model dataset toward answering these questions.

As shown above, the simulated (Fig. 8) and observed (Figs. 1–3) 14–16 December 1987 cyclone possessed a well-defined upper-level front during much of its lifetime. This upper-level front extended from the tropopause into the lower troposphere, where it became juxtaposed over the surface-based cold and warm fronts. In fact, it appeared that the upper-level front supplied what would traditionally be considered the elevated cold front of the occlusion (see Schultz and Mass 1993).

In order to relate the evolution of the upper-level and lower-tropospheric baroclinic zones, (model) temperature fields at four levels (1000, 800, 600, and 400 mb) and times (3, 12, 24, and 33 h) are displayed in Fig. 23. From these analyses it appears that:

1) the upper-level front strengthened through hour 24 of the simulation and subsequently weakened rapidly;
2) the upper front was a smaller-scale feature than the lower-tropospheric baroclinic zone and was closely associated with the upper-level short wave (cf. Fig. 8);
3) the upper front was apparent at the earliest time (3 h). At this time it was relatively distinct from the surface trough and the most intense portion of the lower-tropospheric baroclinic zone. As the system amplified, the upper front and the lower-tropospheric baroclinic zone became more closely aligned.

The separate natures of upper- and lower-tropospheric baroclinic zones and the tendency for the former to catch up with the latter have been noted in several earlier studies. For example, on the basis of case studies, Palmén and Newton (1969) noted that "the upper-tropospheric frontal layer eventually moved forward and combined with lower-tropospheric fronts on the east side of the trough, but only after the cyclones had attained considerable development. Thus, a frontal layer extending through the entire troposphere is, at least in some cases, a characteristic acquired by a cyclone during, rather than prior to, its development."

Additional insight into the origin of the upper-level front during the December 1987 event can be derived from a further examination of the 21-h backward trajectories ending at 24 h shown in Fig. 16. At the $\sigma = 0.395$ level (~450 mb), where the upper front was relatively narrow and well defined, the trajectories ending on the cold side of the upper front (79–81, 79 not shown) originated in the lower troposphere to the south and rose rapidly to their final positions. In contrast, the trajectories within the upper-level front (82–84) started farther to the north and experienced significant subsidence during their westward advance, with the greatest sinking on the warm side of the front (trajectory 84 descended from 229 to 446 mb). The trajectories ending on the warm side of the front (85–88, 86–88 not shown) began in close proximity (to the south of trajectory 84’s initial position) and experienced net descent. It appears that vertical displacement, rather than simple horizontal confluence, dominated the thermal pattern across the upper front at this level; air parcels ending on the cold side of the front (79–81) began farther to the south than those on the warm side of this feature (85–87).

At the $\sigma = 0.595$ level (~630 mb), moist, rapidly rising trajectories (55 and 56) were again found (Fig. 16d) on the cold side of the upper front. Three of the trajectories ending within the upper frontal zone experienced appreciable net descent (i.e., trajectories 57–59, 57 not shown), with the warmest (59) associated with the greatest subsidence (from 331 to 623 mb). In contrast, the fourth trajectory (60), ending on the warm boundary of the upper-level front, rose rapidly from deep within the warm air near the surface (966 mb). Although vertical motions still appear to be important for supporting the horizontal temperature gradient, the confluent contribution is far more apparent at this level, with the trajectories terminating on the cold side (55 and 56) originating far to the north of and at similar levels to the air parcels on the warm side (61–63, 62 and 63 not shown).

At $\sigma = 0.795$ (~800 mb), the confluence of trajectories from different source regions appeared to be the main source of baroclinicity (Fig. 16c).

It is of interest to note that little, if any, cold advection was evident in the region of upper-level frontogenesis (Fig. 8). Keyser (1986) and Keyser and Shapiro (1986) note that two mechanisms—1) confluence accompanied by cold advection and 2) alongstream adjustment to curvature variations—might explain the differential subsidence pattern contributing to upper-level frontogenesis and that the relative importance of these two mechanisms during three-dimensional baroclinic development is still unknown. This modeling study suggests that the cold-advection mechanism was of relatively minor importance during the December 1987 event. A similar finding was discussed in Sanders et al. (1991) for an upper-level front during October 1963.

In summary, during the early part of the simulation, the upper and lower baroclinic zones were relatively distinct. The two baroclinic zones came together as the short wave (and associated upper-level baroclinicity) overtook the low-level baroclinic zone; by 24 h into the simulation, these two features were closely associated and interrelated, with one melding into the other. The upper- and lower-level fronts were also of distinctly different alongfront scales: the upper front was more
Fig. 23. Model temperature (°C) analyses at 3, 12, 24, and 33 h into the simulation for four pressure levels (1000, 800, 600, and 400 mb). Temperatures have a contour interval of 2.5 °C.
FIG. 23. (Continued)
limited in dimension and on the scale of the upper short wave, while the low-level baroclinic zone extended many thousands of kilometers.

Although one baroclinic zone stretched from the surface to the tropopause at 24 h, the origin of this zone varied by elevation. In the lower troposphere, the frontal zone was forced predominantly by quasi-horizontal confluent motions, while in the upper troposphere, differential vertical motions dominated. In the midtroposphere, both forcing mechanisms were apparent. These findings are consistent with earlier observational (Reed and Sanders 1953; Reed 1955; Newton 1954; Reed and Danielsen 1959; Bosart 1970; Shapiro 1970) and theoretical (Shapiro 1981, 1983; Shapiro and Kennedy 1981; Keyser and Pecnick 1985) studies, which have also suggested the dominance of differential subsiding motions in upper-level baroclinic zone development.

d. Comparison of model frontal evolution and the fractured front-T-bone conceptual model of cyclone structural development—The relationship between frontogenesis and thermal structures at the surface

Recent observational studies of marine cyclogenesis using aircraft, dropsondes, and other new observing technologies (Neiman et al. 1990; Shapiro and Keyser 1990), as well as semigeostrophic (Schrä 1989) and primitive equation model simulations (Donall et al. 1991), have suggested cyclone structures and evolutions that differ substantially from the classic Norwegian conceptual model. In many of these studies, the cold front projects northward toward the warm front at nearly a right angle—a pattern often termed the “T-bone”—with the cold front appearing to weaken (or “fracture”) as it approaches the warm front. An important deviation from the Norwegian model is the lack of an occluded front; rather, the cold front appears to propagate normal to the warm front and never catches up with the warm front. In the new model, a back-bent warm front is analyzed to extend through the low, not unlike a bent-back or retrograde occlusion (Bergeron 1937; Petterssen 1956), and possesses a larger horizontal temperature gradient than the cold front. Another important element of the new model is nontopographic seclusion, whereby cold air circling around the low center surrounds and cuts off a core of warmer frontal-zone air.

The model simulation of the December 1987 cyclone, as well as the observed storm itself, suggests both consistencies and contradictions with the new T-bone conceptual model.

- As displayed in the surface temperature plots (e.g., Fig. 12), the northern section of the surface cold front does not significantly fracture or weaken (as in the T-bone model). Unlike the Shapiro-Keyser model, the warm front does not possess an appreciably larger horizontal temperature gradient than the cold front. The surface frontogenesis field does weaken, however, along the northern cold front, and a single area of frontogenesis supports both the warm and occluded fronts.
- Above the surface, the model temperature fields appear more similar to those of the T-bone model. For example, the storm’s 800-mb thermal field at 24 h (Fig. 23) suggests the back-bent warm front and the weaker cold front of the T-bone model.
- In the lower troposphere, the cold front appears to catch up with the warm front as in the classical occlusion process [Fig. 12; see also Schultz and Mass (1993)]; in the T-bone model, such catch-up does not occur.
- There is little evidence of nontopographic seclusion of warm air near the surface during the 36-h simulation. There is, however, a hint of this process at 800 mb at 33 h into the simulation (Fig. 23).

All of the observed and simulated cases used in the development of the Shapiro-Keyser conceptual model were marine cyclogenesis events. In contrast, the December 1987 event and most of the case studies described in the literature (and used in synoptic classes) are continental cases, all of which are structurally different from the Shapiro-Keyser model (e.g., they possess nonfractured cold fronts at the surface) and are more reminiscent of the classical model. In light of these differences between land and ocean events, the natural question is whether 1) one basic process of cyclone evolution is being modified at low levels by differing lower boundary processes over land and water or 2) there is an essential difference in the nature of cyclogenesis over the two regions. A possible source of 2) would be if there are different large-scale flow regimes over the two regions and if such variability leads to differing cyclone structures (Thorncroft and Hoskins 1990; Davies et al. 1991).

Insight into the origins of the structural differences between continental and oceanic storms can be gained by examining the relationship between cyclone frontogenesis and thermal structures. Although thermal structures differ significantly (fractured versus nonfractured cold fronts, occluded fronts versus back-bent warm fronts), the distribution of Lagrangian frontogenesis does not appear to vary greatly. For example, as noted by Takayabu (1986), Hines and Mechoos (1992, personal communication), Schär (1989), and explicitly shown in this paper (Fig. 15), virtually every cyclone simulation shows a fracturing of the cold and warm frontogenesis, with the clear dominance of the later. Davies-Jones (1985) and Doswell (1985) have shown that this fracturing is a basic feature of the interaction of vortices with preexisting temperature gradients. In all of the aforementioned simulations, there is no distinction between the frontogenesis supporting the warm and occluded fronts; that is, one area of frontogenesis, of common origin, clearly supports both.
Considering the similarities among the basic frontogenesis fields of varied simulations over land and water, why should the resulting thermal structures vary so noticeably among continental and oceanic cyclones? Why are continental cold fronts rarely fractured even though the frontogenesis evinces a substantial weakening along the northern part of the cold front? A possible answer lies in the profound difference in the trajectories of air parcels within the cold- and warm-frontal zones. As shown by Schär and Wernli (1992), Takayabu (1986), and in section 5e of this paper, a warm front is essentially a three-dimensional entity in which new air parcels are continuously and rapidly being ingested into the frontal zone and expelled. Air parcels destined to enter the warm-frontal zone begin east of the system within regions of weak horizontal temperature gradient. With a continuous confluent influx of new air parcels from the east, Lagrangian frontogenesis must be large enough to maintain a large temperature gradient within the warm-frontal zone. In contrast, the air parcels ending within the cold-frontal zone usually begin within the baroclinic zone. For example, in section 5 it was noted that the parcels ending within the northern section of the cold front began in the baroclinic zone to the northeast of the low center and swung around the low. Since air parcels ending within the cold-frontal zone also begin within a baroclinic zone, large Lagrangian frontogenesis is not needed to maintain frontal temperature gradients. It is for this reason that over land, the cold front can remain unfractured even though frontogenesis along its northern section is weak. Furthermore, enhanced friction over land contributes some frontogenetical forcing that helps maintain continental cold fronts near the surface.

Over the ocean, the situation is different. The lower troposphere over the ocean, especially over the western ocean basins during winter, is generally a frontalitical environment, with large fluxes of sensible heat and water vapor rapidly modifying cold continental air masses as they pass over the warmer water. Thus, temperature gradients within the area of weak frontogenesis along the northern cold front can attenuate rapidly, producing a fractured cold-frontal structure. Lesser surface drag over the ocean reduces frictional convergence and associated frontogenesis for both warm and cold fronts; however, for the case of the warm front, reduced frictional frontogenesis is compensated by the stronger winds and thus greater confluent frontogenesis that accompanies the lower drag. It is this confluence that supports both the traditional warm front as well as the back-bent warm front—occlusion observed in oceanic storms. Stronger winds also increase the length of the back-bent warm front because they enhance the advection of the frontal zone into the cold frontalitical environment to the rear of the cyclone.

Reduced surface drag over the ocean implies larger low-level wind velocities and longer air-parcel trajectories during a given length of time. Such longer trajectories might help explain the apparent greater oceanic frequency of occlusions, in which cold air is able to completely encircle the low, trapping warmer frontal-zone air in the center. Consistent with this speculation is the observation that just above the boundary layer, where friction is less of a factor (e.g., 800 mb), the model thermal fields in the December 1987 land case suggest the occlusion process (e.g., Fig. 23). Further confirmation of the importance of varying surface drag in differentiating continental and oceanic storms has been provided by a recent simulation by Hines and Mechoso (1992, personal communication) in which a 21-layer, primitive equation model produced cyclones with weak warm fronts and unfractured cold fronts using surface drag appropriate for land; reducing the surface drag to oceanic values resulted in structures reminiscent of the Shapiro–Keyser model.

Another question deals with the apparent lack of occluded fronts and their replacement with (back-bent) warm fronts in some oceanic cyclones (Shapiro and Keyser 1990). In the December 1987 land cyclone, the occluded front is a highly frontogenetical feature and it is impossible to separate or differentiate the frontogenesis along the warm and occluded fronts. The unitary nature of occluded–warm frontogenesis has been demonstrated in many other simulations (e.g., Takayabu 1986; Schär and Wernli 1992) and is the result of the horizontal confluence of westward-moving air with differing initial temperatures (Figs. 16, 18 of this paper; Kuo et al. 1992). As a result, one is compelled to ask whether there is an essential difference between continental occluded fronts, such as the one found in the December 1987 cyclone, and the bent-back warm fronts in oceanic cyclones described in Shapiro and Keyser (1990). As noted above, many of the differences between land and ocean structures can be explained by differing boundary-layer processes. Over the ocean, the cold front is weakened by a frontalitical lower boundary and lessened frictional convergence, while at the same time, the (back-bent) warm front is enhanced by the higher wind speeds. Aloft the Shapiro–Keyser back-bent warm front lacks the elevated cold-frontal zone of a canonical warm-type occlusion. But it is important to note that the elevated "cold" front of the December 1987 occlusion was not the result of a surface-based cold front riding aloft over the warm front (as it would be in a classical Norwegian model occlusion). Instead it was the downward extension of an upper-level front (associated with a short wave–jet streak aloft) that only extended over a portion of the lower-tropospheric occluded front.

7. Summary

Using both observational data and the output from the PSU–NCAR Mesoscale Model, this paper describes the evolution of the temperature and humidity structures of an intense cyclone that evolved over the eastern
half of the United States on 14–16 December 1987. The storm simulation was generally quite successful, capturing the intensity and movement of the cyclone, as well as most of its structural characteristics. Problems with the simulation include modest position and intensity errors during the initial and later portions of the simulation, as well as slow cold-frontal speed and the resulting delay of occlusion. Although the precipitation pattern was generally quite realistic, the amounts were overestimated in some areas by as much as 50%–100%.

Vertical and horizontal cross sections through the simulated cyclone suggest both similarities and differences with the structures described in the classic Norwegian cyclone model. For example, the occlusion process at the surface appears similar to the classical paradigm, with the cold front catching up with the warm front. Unlike the Bergen school model, a warm-type occlusion structure formed as an upper baroclinic zone merged with the surface-based occluded front, which appeared to form as the surface-based cold front caught up with the surface-based warm front.

Initially, the frontogenesis associated with the cold and warm fronts was continuous, but as the system evolved, a break in the frontogenesis along the northern section of the cold front developed. A single frontogenesis feature appears to support both the warm and occluded fronts.

Trajectories derived from the cyclone simulation revealed that many of the structural elements of the storm can be explained by the differing air-parcel trajectories across these features. Although most trajectories can be meaningfully grouped into a limited number of families, they cannot be presented accurately in terms of only two or three conveyor belts or airstreams.

In this cyclone, as in many others, a quasi-linear region of convective activity was observed in advance of the surface cold front. The convective line was associated with an upper-level humidity front, which originated from the confluence of descended, mid-, and upper-tropospheric trajectories and saturated, rising warm-sector trajectories. This configuration resulted in a zone of potential instability, which may have been released by rising motion above the warm sector.

During the early part of the simulation, the upper and lower baroclinic zones were relatively distinct, with the upper baroclinic zone associated with an upper-level short wave–jet streak. The two baroclinic zones came together as the short wave overtook the low-level baroclinic zone. The lower-tropospheric baroclinic zone was forced predominantly by quasi-horizontal confluent motions, while differential vertical motions dominated for the upper-troposphere baroclinic zone. The upper baroclinic zone was more limited in horizontal dimension, being on the scale of the upper short wave.

The model simulation of the December 1987 cyclone, as well as the observed storm itself, suggests both similarities and differences with the T-bone conceptual model of Shapiro and Keyser (1990). Unlike the Shapiro–Keyser model, thermal seclusion and significant fracturing of the cold front were absent at the surface; however, aloft the simulation and the Shapiro–Keyser model are more consistent. It is speculated that the differences in the structures of the December 1987 cyclone and the Shapiro–Keyser model are attributable to the differing surface properties of land and ocean.

Acknowledgments. This research was funded by the National Science Foundation under Grant ATM-8912472. We gratefully acknowledge the contributions of Dr. Ying-Hwa Kuo and Dr. Richard Reed for making possible our acquisition of the model data used in this research. We also thank Dr. George Grelle, of NCAR’s Microscale and Mesoscale Meteorology Division, who ran the MM4 simulations, and Mr. Mark Albright for assistance in computing some of the model trajectories and for many useful suggestions. Yea-Ching Tung and Mark Stoeilinga wrote the trajectory-plotting software. Two anonymous reviewers and Dr. Fred Sanders provided numerous comments that materially improved the manuscript. The mesoscale model was run at the Scientific Computing Division at the National Center for Atmospheric Research, which is supported by the National Science Foundation.

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