Assessment of Slantwise Convection in ERICA Cyclones

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ABSTRACT

Atmospheric stability properties for cumulus and slantwise convection in oceanic midlatitude cyclones are analyzed using dropsonde observations from the Experiment on Rapidly Intensifying Cyclones over the Atlantic (ERICA). Vertical cross sections perpendicular to the low-level wind shear are selected in the frontal regions for four ERICA storms. To assess the stability properties for conditional symmetric instability (CSI), a sounding analysis is carried out along surfaces of constant absolute angular momentum $M$. The buoyancy of the parcel along the slanted $M$ surface is determined, both with and without the water loading effect. Our analysis suggests that a systematic bias toward overestimation of slantwise instability occurs when the loading effect is neglected.

The major finding of our analysis is that the lower-tropospheric air on the warm side of the warm-frontal zone is stable or neutral with respect to vertical cumulus convection but unstable for slantwise convection. Convective instability, however, is found in the warm sector near the surface low of explosive cyclones during their period of most rapid growth. Our analysis shows that conditional slantwise instability, throughout a deep layer, can occur even in a slowly developing cyclone. Observed precipitation events were consistent with the occurrence of slantwise and cumulus convection.

1. Introduction

In recent years, much attention has been given to the role of conditional symmetric instability (CSI, also termed slantwise convection) in the development of bands of slanted upward motion in a baroclinic flow. Kuo (1954), Stone (1966), and Ooyama (1966) identified that symmetric instability in a baroclinic circular vortex arises from an imbalance of buoyancy forces in the vertical and centrifugal forces in the horizontal. Since the vortex flow is zonally symmetric, the term symmetric instability is used to distinguish it from the classical azimuthally modulated baroclinic instability.

Emanuel (1979) showed from linear stability analysis that the horizontal scale for dry symmetric unstable motion ranges from about 50 to 500 km and have a time scale of a few hours. Clearly, both the spatial and temporal scales are much smaller than those of baroclinic processes such as cyclogenesis, establishing symmetric instability as a mesoscale process.

Bennetts and Hoskins (1979) first suggested that CSI may be important in the formation of precipitation bands. They showed that slantwise overturning can occur when the equivalent potential vorticity becomes negative in a geostrophically balanced unidirectional airflow. Alternatively, the necessary condition for CSI is that the slope of the absolute angular momentum surfaces is shallower than that of the surfaces of equivalent potential temperature. Emanuel (1983) extended this idea and showed how slantwise instability can be deduced using data from a synoptic network of sounding observations. Concurrent and subsequent studies (Parsons and Hobbs 1983; Seltzer et al. 1985; Sanders 1986; Wolfsberg et al. 1986; Moore and Blakeley 1988; and others) have shown that CSI does occur locally in midlatitude weather systems and there is considerable evidence that it may be responsible for the occurrence of banded structures in the precipitation field.

In a study of an isolated precipitation band, however, Sanders and Bosart (1985) found that the band persisted for several hours in an environment neutrally stable to slantwise convection. Emanuel (1985) and Thorpe and Emanuel (1985) offered an explanation by demonstrating that frontogenesis is especially rapid in a slantwise neutral environment and that very narrow slanted updrafts can be generated. Because slantwise convection is often associated with frontal development, it is difficult to distinguish separately the two mechanisms.

Very recently, additional interest in CSI was spawned from its possible role in explosive cyclogenesis [see Sanders and Gyakum (1980) for a definition]. It is
known that a substantial source of energy for cyclogenesis over warm ocean waters is latent heat release through condensation in clouds (e.g., Gyakum 1983). Numerical modeling results of maritime storms often exhibit regions of upright or slantwise instability in the warm-frontal zone (Kuo and Reed 1988; Mailhott and Chouinard 1989; Hedley and Yau 1991). What role upright and slantwise convection plays in the rapid spinup of these systems remains an important question.

The above discussion has shown that the distribution of the stability properties of the atmosphere to both upright and slantwise convection has important implications for the organization of precipitation, frontogenesis, and explosive marine cyclogenesis in mid-latitudes. In the past, it has been extremely difficult to document unambiguously the occurrence of slantwise convection in marine cyclones. Observations of mesoscale structures are difficult to obtain, particularly when the phenomena occur over the ocean where upper air soundings are not taken routinely. However, observations from recent field projects such as the Genesis of Atlantic Lows Experiment (GALE), the Canadian Atlantic Storms Program (CASP), the OCEAN STORMS study of polar lows, and the Experiment on Rapidly Intensifying Cyclones over the Atlantic (ERICA) are providing unique opportunities for examining mesoscale structures in coastal and oceanic storms. Some results are already in hand. Emanuel (1988) has analyzed aircraft soundings made during GALE and showed that the air in the warm sector of winter cyclones is stable to vertical convection but close to conditional slantwise neutrality. Apparently, an adjustment of the atmosphere was taking place to bring it toward its slantwise convectively neutral state. Radiosondes observations during CASP (Donaldson and Stewart 1989; Reuter and Yau 1990) have indicated several cases where the atmosphere contains slightly unstable regions for CSI, usually within air layers that have very strong vertical shear. Bond and Shapiro (1991) have analyzed dropwindsonde data in a polar-low storm sampled during the OCEAN STORMS experiment. They found that the lower troposphere in the warm sector was close to neutral with respect to slantwise convection.

The purpose of this paper is to assess the stability field for both (upright) cumulus convection and slantwise convection in the frontal zones associated with developing oceanic cyclones. Measurements from the Experiment on Rapidly Intensifying Cyclones over the Atlantic (ERICA) serve as the database for our analysis. Of the eight intensive observing periods surveyed, four have been found to have sufficient sounding data to allow an assessment of CSI. The organization of the precipitation field based on available radar observations is also examined to establish a possible link with slantwise convection. Our study is novel in two major aspects.

1) The sounding data from LeSondes released from research aircraft have a horizontal resolution of 120–150 km, considerably finer than that used in previous studies.

2) Our stability analysis includes the effects of water loading in estimating the buoyancy of the cloudy air (Xu and Emanuel 1989). This avoids the overestimation of CSI due to the neglect of the water drag.

The organization of the paper is as follows. The theoretical and computational aspects of CSI are presented in the next section. Background information of the ERICA field study are discussed in section 3, followed by a discussion of the way in which cross sections were constructed in section 4. The results of the CSI stability analysis are presented in sections 5–8, each dealing with one storm. The final section contains the conclusions and a discussion of the possible effects of the stability field on cyclogenesis.

2. Assessment of conditional symmetric instability

To assess CSI, the standard analysis technique (Emanuel 1983) is extended to include the effects of environmental humidity and condensed water on the buoyancy of cloudy air. To provide a suitable framework, the theory of CSI is briefly outlined first. Consider a baroclinic atmosphere with a geostrophically balanced flow. The equivalent potential vorticity \( q_e \) is given by

\[
q_e = \frac{1}{\rho} \left[ \frac{\partial v}{\partial x} + f \frac{\partial u}{\partial y} \right] \left[ \frac{\partial \theta_e}{\partial z} - \frac{\partial v}{\partial z} \frac{\partial \theta_e}{\partial x} + \frac{\partial u}{\partial z} \frac{\partial \theta_e}{\partial y} \right],
\]

where \( u \) and \( v \) are the \( x \) and \( y \) components of the wind, \( \theta_e \) is the equivalent potential temperature, \( f \) is the Coriolis parameter, and \( \rho \) is the air density. The \( x \) axis is directed perpendicular to the main thermal wind shear. The theory of CSI is developed for a two-dimensional flow with no variation along the \( y \) direction. Under such a condition, the expression for \( q_e \) reduces to

\[
q_e = \frac{1}{\rho} \left[ \frac{\partial v}{\partial x} + f \frac{\partial u}{\partial y} \right] \left[ \frac{\partial \theta_e}{\partial z} - \frac{\partial v}{\partial z} \frac{\partial \theta_e}{\partial x} + \frac{\partial u}{\partial z} \frac{\partial \theta_e}{\partial y} \right].
\]

Bennetts and Hoskins (1979) showed that CSI in saturated air occurs whenever the value of \( q_e \) becomes negative. An alternative statement is that the saturated air becomes unstable to symmetric disturbances when the slope of an \( M \) surface is less than that of a \( \theta_e \) surface, where \( M \) is the absolute angular momentum defined by \( M = v + fx \). Both these criteria deal with the local instability for an infinitesimal displacement of a tube of air infinitely long in the \( y \) direction.

For moist but unsaturated air, the presence of negative \( q_e \) values is only a necessary condition for instability, and a more thorough analysis is required to determine a sufficient condition. In particular, one should consider whether a finite displacement of the tube of
air could result in saturation. Since the forces acting on the tube are conservative, the energy changes depend only on the starting and end points, while the exact path of displacement is immaterial. For assessing CSI, it is convenient to assume the displacement to follow an $M$ surface. The extent of slantwise instability then depends only on the buoyancy of the tube of air, measured by the difference in virtual temperature between that of the cloudy and the ambient air. The virtual temperature $T_v$ of the ambient air is given by

$$T_v = T \frac{1 + 1.61r}{1 + r},$$

where $T$ and $r$ are the temperature and vapor mixing ratio of the ambient air. The virtual temperature of the tube of cloudy air depends on whether the condensation process is assumed to be pseudoadiabatic or reversible adiabatic. In the pseudoadiabatic case, the condensed water is assumed to precipitate out instantaneously, whereas in the reversible moist-adiabatic process, the parcel retains all condensed water. Thus, for the reversible adiabatic process, the effective density of the cloudy air is increased by the loading effect of the condensed water. The expressions for the virtual temperatures (Xu and Emanuel 1989) are

$$T_v(\text{reversible adiabatic}) = T_p \frac{1 + 1.61r^*(T_p)}{1 + r^*(T_p) + r_c},$$

$$T_v(\text{pseudoadiabatic}) = T_p \frac{1 + 1.61r^*(T_p)}{1 + r^*(T_p)},$$

where $T_p$ is the temperature of a displaced parcel, $r^*$ is the saturation mixing ratio, and $r_c$ is the adiabatic cloud-water mixing ratio. The parcel temperature $T_p$ is derived from $\theta_e$, which is computed using Bolton's (1980) formula. In calculating $\theta_e$ and $T_p$, differentiation between the reversible and pseudoadiabatic process is unnecessary because the respective values differ only marginally (Iribarne and Godson 1981; p. 143), well within the error margin of the temperature measurements.

3. Experiment of Rapidly Intensifying Cyclones (ERICA)

The ERICA field study was designed to improve the understanding of the physical mechanisms that cause rapid development of winter cyclones over the ocean. Major points of investigation included the roles of mesoscale processes such as destabilization of the lower atmosphere, cyclogenetic forcing by surface fluxes, and cumulus–slantwise convection. Hadlock and Kreitzberg (1988) provided an overview of the project and summarized the major hypotheses that were tested. Their report also includes a map of the ERICA data acquisition network and an overview of the various observation platforms that were utilized during the field phase.

The ERICA field phase was conducted from 1 December 1988 through late February 1989. A total of eight intensive observing periods (IOPs) were called during which major oceanic storms occurred (Harnett et al. 1989). Of the eight storms, four were very rapidly deepening cyclones: IOP 1 [18 hPa (9 h$^{-1}$)]; IOP 2 [32 hPa (18 h$^{-1}$)]; IOP 4 [60 hPa (24 h$^{-1}$)]; and IOP 5 [36 hPa (18 h$^{-1}$)]. Two of the storms did not result in a closed surface isobars, but they revealed frontal structures (IOP 6 and IOP 7).

Several instrumented aircraft were deployed to sample the mesoscale structure of the cyclones. They had the major advantage of being mobile and were not limited by a particular storm track. The aircraft flights were carefully scheduled to meet the central needs of the field study. The first mission was essential to document the preexisting environmental baroclinicity, static stability, and moisture, particularly in the airmass that was later ingested into the storm at low levels during its rapid intensification phase. The second aircraft usually took off 12 h after the first. Its mission was to sample the wind during the early stages of cyclogenesis and to document the evolution of the surface pressure center. The third mission was undertaken again by the first aircraft and took place an additional 12 h later. This staggering of 12 h between takeoffs made it possible to sample data through the complete life cycle of a cyclone. The typical flying time of a mission was 9 h, of which one-third was spent in cruising toward and back from the storm. A summary of the synoptic conditions and the flight missions for each IOP are given by Harnett et al. (1989).

This study is based mainly on the measurements made by the LeSondes released from the two WP-3D aircraft. LeSonde is an aircraft-deployed lightweight dropwindsonde package that uses Loran C navigation signals to measure the dropsonde's trajectory, from which horizontal winds are computed. The sonde falls at about 9 m s$^{-1}$ near the surface and about 12 m s$^{-1}$ at 50 kPa. A sounding from the 50-kPa cruising altitude takes about 20 min. Wind measurement accuracies are about 1 m s$^{-1}$, slightly worse in highly convective conditions. The LeSondes also recorded temperature and dewpoint with error margins of about 0.5$^\circ$ and 1.5$^\circ$C, respectively.

4. Strategy for constructing vertical cross sections

The source of our LeSonde sounding data is the 1-kPa resolution data prepared by the ERICA Data Center at Drexel University. About 400 soundings were recorded. The soundings were carefully scrutinized for data suitable to construct vertical cross sections perpendicular to a flow pattern that was locally two-dimensional. A selection criterion is that the alongfront
scale must be at least five times greater than the crossfront scale. In addition, the LeSonde data for the cross sections also had to be sampled within a 90-min time lapse in order to portray a "snapshot" structure of the atmosphere. With these criteria in mind, six cases were selected. Their IOP numbers, together with the dates and times of release of the LeSonde (in UTC), are listed in Table 1.

The direction of the y axis in our analysis was chosen in the direction of the vertical shear of the observed wind between 90 and 70 kPa. Above 70 kPa, the shear was usually small, and beneath 90 kPa, boundary-layer friction causes the wind to deviate from its geostrophically balanced state. The positions of the dropsonde locations are projected onto the x axis. The M values are approximated by using \( M = fx + u \), where \( x \) is the projected dropsonde location and \( v \) is the \( y \) component of the observed wind, which was always close to the thermal wind component computed from the temperature measurements.

Reuter and Yau (1990) have given a comparison of intersounding distances in various studies on assessing CSI. The typical spacing between adjacent soundings was about 250–500 km, barely adequate to resolve the mesoscale. In the ERICA LeSonde data, however, the spacing between adjacent drops was about 120–150 km, a significant improvement over previous studies. Linear interpolation in the horizontal was used to obtain \( M \) and \( \theta_e \) values on a grid with horizontal spacing of 100 m and a vertical resolution of 1 kPa. From these "gridded" data, the \( q_r \) field is computed using a finite-difference approximation. With error margins in temperature, dewpoint, and wind speed being within ±0.5°C, ±1.5°C, and ±1 m s⁻¹, respectively, the uncertainty in \( q_r \) values is estimated to be within ±30% of its actual value.

5. Results for the frontal zone of 4 January 1989

The cyclone of 3–5 January 1989 (ERICA IOP 4) may have been the deepest extratropical cyclone observed in this century south of 40°N latitude. The cyclone formed and intensified over the warm Gulf Stream waters east of 75°W longitude. The central surface pressure decreased by 65 hPa to 935 hPa in 24 h during its rapid-development phase from 0000 UTC 4 January to 0000 UTC 5 January (Shapiro and Keyser 1990). The cyclogenesis occurred along the remnants of a weak cold front just south of the SST gradient. The warm flow to the east of the surface low created a tight baroclinic zone. When the upper-tropospheric wave disturbance caught up with the surface disturbance, rapid cyclogenesis occurred. The intensifying low moved south of and parallel to the SST front.

To sample the developing cyclone, five aircraft missions were conducted (Harnett et al. 1990). The aim of the first mission was to observe the structure of the midtroposphere at the very early stage of the storm's development. Sanders' (personal communication) analysis at 0600 UTC (Fig. 1) shows a baroclinic zone lying immediately to the north of the surface low.

Measurements from four consecutive drops were used to construct the \( M, \theta_e \), and \( q_r \) surfaces projected onto a vertical plane oriented perpendicular to the vertical shear in the 90–70-kPa layer. On the synoptic scale, the low-level flow pattern exhibited three-dimensional structures. However, throughout the section AB measuring 600 km in length, the variation of winds in the x direction was much larger than that in the y direction. Hence, locally the mesoscale flow was nearly two-dimensional.

The surface low is located at about \( x = 20 \) km in Fig. 2. Here \( \theta_e \) decreases in the vertical in the boundary layer \( (x > 100 \) km, \( p > 90 \) kPa) and above the surface low \( (x < 100 \) km, 95 kPa < \( p < 80 \) kPa). Thus, when water-loading effects are neglected, the air is potentially unstable to upright convection above and just north of the surface low center. For this particular case, regions of convective instability are mostly contained within regions that are potentially unstable for CSI, as is illustrated by the shallower slopes of the constant-\( M \) surfaces relative to the \( \theta_e \) surfaces. Alternately, the potential for CSI is marked by negative values in the \( q_r \) field. The bottom panel in Fig. 2 shows clearly that the potential instability for slantwise convection decreased northward and the air was near neutral stability \( (q_r = 0) \) for slanted ascent below 90 kPa for \( x > 300 \) km. Since the time scale for convective instability (several minutes) is much shorter than that for CSI (a few hours), slantwise convection would become important after upright convection has taken place.

The effects of humidity and water loading on the buoyancy are included to more accurately assess conditional instabilities. Figure 3 illustrates the ambient \( T_r \) and \( T_d \) soundings along selected different \( M \) surfaces. Also plotted are the \( T_r \) profile of the parcel lifted from the lifting condensation level (LCL) with and without the inclusion of the drag effect of the condensed water. For comparison purposes (and for the assessment of conditional upright instability), the corresponding curves for upright convection passing through the LCL along the \( M \) surfaces are included. Figures 3b, 3d, 3f, and 3h indicate that when water loading is taken into account, the atmosphere is conditionally neutral or stable to upright convection. However, CSI is evident.

| Table 1. Cases of analysis with time (UTC) of releasing the LeSonves. |  
|--------------------------|-----------------|-----------------|-----------------|-----------------|-----------------| 
| IOP 2 13 December 1988:  | 1400            | 1420            | 1443            | 1506            | 1526            | 
| IOP 2 14 December 1988:  | 2051            | 2032            | 2140            | 2215            | 2220            | 
| IOP 4 4 January 1989:    | 0045            | 0523            | 0604            | 0620            | 0630            | 
| IOP 7 13 February 1989:  | 0109            | 0128            | 0203            | 0220            | 0225            | 
| IOP 8 24 February 1989:  | 1737            | 1754            | 1810            | 1840            | 1850            | 
| IOP 8 24 February 1989:  | 2028            | 2044            | 2101            | 2117            | 2120            |
FIG. 1. Analysis of sea level isobars (solid) at 4-hPa intervals and of surface isotherms (dashed) at 5°C at 0600 UTC 4 January 1989. The isopleths were redrawn from Sanders' (1990) analysis. The solid dots show the four selected LeSondes deployment positions used for data analysis. The segment AB indicates the projection line for Fig. 2.

particularly along the \( M = 0 \) m s\(^{-1}\) surface (Fig. 3c). Regions of CSI are confined to the lower levels north of the storm center and below 76 kPa along \( M = 20 \) m s\(^{-1}\).

FIG. 2. Vertical cross section constructed from LeSondes's data at about 0500 UTC 4 January 1989, projected along AB in Fig. 1. The top panel shows contours of \( \theta_e \) (solid) and \( M \) (dashed) contoured at 2-K and 10 m s\(^{-1}\) intervals. The arrows indicate the positions of the LeSondes' falls. Shading denotes relative humidity values above 80%. The bottom panel shows contours of \( q_v \) at intervals of 0.05 PVU (or \( 5 \times 10^{-6} \) K m s\(^{-1}\) kg\(^{-1}\)).

FIG. 3. Skew T–logp diagrams showing the distribution of virtual temperature and dewpoint along selected paths at about 0500 UTC 4 January 1989. The dashed curves show the \( T_v \) of a saturated parcel being lifted from the lifting condensation level (LCL), both with and without the drag effect of the condensed water on the density in the definition of \( T_v \). (a) Slanted sounding along \( M = -10 \) m s\(^{-1}\) surface. (b) Vertical sounding passing through \( M = -10 \) m s\(^{-1}\) at LCL. (c) Slanted sounding along \( M = 0 \) m s\(^{-1}\) surface. (d) Vertical sounding passing through \( M = 0 \) m s\(^{-1}\) at LCL. (e) Slanted sounding along \( M = 10 \) m s\(^{-1}\) surface. (f) Vertical sounding passing through \( M = 10 \) m s\(^{-1}\) at LCL. (g) Slanted sounding along \( M = 20 \) m s\(^{-1}\) surface. (h) Vertical sounding passing through \( M = 20 \) m s\(^{-1}\) at LCL.

In the flight report, it was noted that heavy precipitating rainbands with lightning activities were encountered (Harnett et al. 1990). Radar reflectivity measurements from the lower fuselage radar were as
high as 50 dBZ during 0450-0510 UTC. The intense rainfall rates are likely the manifestation of the effects of CSI depicted in the soundings.

As cyclogenesis continues, a deep layer of convective instability develops in the vicinity of the surface low. Shapiro and Keyser (1990, p. 185) presented an analysis of the sea level pressure and temperature field for the cyclone at 1700 UTC. At this stage, the storm center was located at 38°N, 63°W, with a surface pressure of 95.2 kPa. A LeSonde sounding about 50 km southeast of this center location indicates conditional instability to upright convection even when the water loading effect is included (Fig. 4). The unstable region extends from the lifting condensation level up to the cruising level at about 50 kPa. Farther to the northeast, the convectively unstable layer was found to be shallower.

6. Results for the frontal zones of 13–14 December 1988

On 13–14 December, rapid cyclogenesis occurred over the eastern Atlantic at latitudes between 35° and 40°N. The central pressure of the storm fell by 18 hPa between 0300 and 0900 UTC on 14 December and reached a minimum of about 960 hPa. Strong winds on the northwest side of the mature cyclone capsized and sunk the deep-sea drilling rig Rowan Gorilla as it was under tow from the oil fields offshore Nova Scotia.

The development of this cyclone was associated with an upper-level jet streak and short-wave trough that moved offshore from Virginia at about 1200 UTC on 13 December. When this feature passed over a complex multicentered surface low pattern, a single center of great intensity developed. The satellite imagery showed a shield of deep cloud with embedded convection (Harnett and Browne 1989). After 0600 UTC on 14 December, a huge comma-shaped cloud mass evolved.

When the cyclone matured, a clear eye at the center of the intense circulation was indicated. It seems that the upper-air trough imposed such an impressive response on surface cyclogenesis, mainly due to the favorable preconditioning of the lower troposphere. The precursor stage at the surface was characterized by a rather weak and unorganized circulation. Numerous mesoscale shallow and deep cloud bands were visible in the satellite imagery.

At 1200 UTC 13 December, a surface low was located at 30°N, 67°W (Fig. 5). A surface trough extended northward in cooler air but was still at the warm edge of a band of strong contrast between the coast and the offshore waters. We focus on the region around 35°N, 68°W, which lies along the track of the surface low center. The cyclone passed over this area 12 h later when it was in its explosive deepening phase. Drifting buoy data showed that the sea surface temperature was about 21°C and the air temperature about 5°–10°C cooler. Significant fluxes of heat into the atmosphere can be expected under such conditions. Also present was a temperature gradient of about 1°C (100 km)−1, yielding a strong thermal shear vector toward the east. Thus, the local pattern of airflow is close to two-dimensional and perpendicular to the segment AB in Fig. 5.

To analyze slantwise stability, data were projected from five LeSondes released at 1400, 1420, 1443, 1506, and 1526 UTC. Surfaces of constant θs show little vertical variation throughout the deep boundary layer up to 80 kPa between x = 200 and x = 300 km (Fig. 6). This marine boundary layer can be considered well mixed and close to neutral with respect to vertical cumulus convection. The layer between 80 and 60 kPa shows a significant wind shear in the M surfaces, particularly in the colder air (i.e., in the northern region). The deep boundary-layer air, in its neutral state to convective instability, has a small potential for slantwise convection. A more substantial potential for CSI is found in the sloping tongue of air in the strong shear regime between 80 and 70 kPa. Note that this region is clearly stable to upright convection.

Figures 7a,b compare the vertical and slantwise ascents (along $M = 15$ m s$^{-1}$). The vertical ascent is close to conditional neutrality up to about 80 kPa and becomes stable thereafter. The slanted ascent, however, indicates conditional slantwise neutrality up to about 68 kPa. The strong wind shear thus has "extended" the marine boundary layer by an additional pressure depth of about 12 kPa. The analysis along the $M = 30$ m s$^{-1}$ surface gives similar results (Figs. 7c,d). The slantwise ascent yields a deeper layer of conditional neutrality compared with that for vertical ascent. In addition, the slanted displacement provides a small enhancement of the potential of upward buoyancy close to the ocean surface. Only a very small amount of kinetic energy is required to raise a parcel from the
surface to its level of free convection, because in the boundary layer, the temperature lapse rate is almost neutral to unsaturated ascent.

Along the flight leg $AB$, the aircraft cruised through convective precipitation (as documented by G. W. Reuter in the cloud physics report for this flight available from the Atmospheric Environment Service). Unfortunately, the lower-fuselage and the tail radars were not operational during the time of analysis.

The surface map based largely on data from low-level aircraft traverses through the cyclone core and on data from surface buoys (see Sanders 1990) depicts the peripheral flow surrounding the warm cyclone center (Fig. 8). Shapiro and Keyser (1990) suggested that this warm bent-back occlusion (or “T-bone structure” in their terminology) is characteristic for intense marine cyclones. At about 2100 UTC, 3 h later, a cross section was sampled passing through this bent-back frontal
zone (Fig. 9). Since the flow has a three-dimensional pattern, \( M \) is not a conserved variable for this case, and thus the theory of CSI is not relevant. Yet, information about the potential for upright convective instability is still evident (Fig. 9). The \( \theta_e \) field reveals that the air above the surface center (\( x = 135 \) km) was neutral for vertical convection up to 70 kPa. The vertical arrangement of the \( M \) surfaces show that there was almost no shear in the \( y \) component of the wind in this region. The warmer air to the south was potentially unstable for cumulus convection since \( \theta_e \) decreases with height.

7. Results for the frontal zone of 12–13 February 1989 (IOP 7)

In spite of strong upper-level forcing and a quasi-stationary low-level frontal zone, cyclogenesis did not occur during IOP 7. Just when the upper-level short-wave trough arrived over the frontal zone, a moderate northwesterly low-level flow developed. This resulted in cold-air advection that counteracted the effect of the positive vorticity advection (PVA) associated with the upper-level flow, and rapid deepening did not ensue.

An opportunity was found for assessing the potential of CSI at about 0200 UTC, when the aircraft crossed over a low-level frontal zone (Fig. 10) and four Le-Sondes were released in a line roughly perpendicular to this frontal zone. The surface conditions as analyzed for 0000 UTC by the Canadian Regional Finite-Element Model showed calm conditions; thus, the 85-kPa chart was chosen to depict the flow. The 2-h lag between the times of map analysis and our stability analysis probably had shifted the trough slightly to the northeast, making the local flow more two-dimensional across the line \( AB \). A cross section of \( \theta_e \) and \( M \) fields is presented in Fig. 11. The strong horizontal gradient in \( \theta_e \), about 7°C (100 km)\(^{-1} \) for the section 0–300 km, causes a significant vertical shearing—indicated by the strong vertical gradients of the \( M \) contours. The atmosphere is stable for upright convection. The warm air, \( x < 300 \) km, is however, potentially unstable for CSI (\( q_e < 0 \)) from the surface up to 80 kPa. The potential unstable air is capped by a deep layer of air close to its state of slantwise neutrality. A small portion of the section located at 70 kPa and \( x > 200 \) km exhibit stable conditions for CSI. The vertical soundings in-
dicated close to neutral conditions in the 90–80-kPa layer and stable conditions above (Fig. 12). Along surface of constant $M = 20 \text{ m s}^{-1}$, however, the air was unstable from cloud base up to the aircraft cruising level (58 kPa).

The potential for slantwise convection indicated above was consistent with flight mission reports that documented the occurrence of intermittent heavy rain. At 0000 UTC (i.e., 2 h prior to our stability analysis), a precipitation band oriented parallel to the low-level shear was intersected. Its width was about 50 km at the freezing level (~2700 m ASL), but its intense convective core aloft was less than 10 km wide.

8. Results for the cyclone of 24–26 February 1989 (IOP 8)

The cyclone track was along a sharp front over the ocean that extended southwest to northeast. There was a marked shift of the wind direction across this occluded front. The air from the south was riding up and over the front, thereby marking a cloud edge that could easily be tracked by aircraft for several hundreds of kilometers. The upper-level forcing for the cyclogenesis was a short-wave trough that traveled through the rear of a long-wave trough axis. By 1200 UTC 24 February, the vorticity maximum associated with the short wave had rounded the base of the long-wave trough and was heading northeastward along the East Coast. Most of the PVA associated with the short-wave trough remained mainly within the cold air spawning a coastal snowstorm. Since little PVA moved sufficiently offshore to reach the stationary low-level front, the cyclone lacked the support to intensify explosively. Its peak deepening rate was estimated at 9 hPa (6 h)~1 during 1500–2100 UTC 24 February.

During this phase of most rapid deepening, the research aircraft traversed the stationary front (from 65° to 68°W along 39.5°N) releasing three LeSondes. The $M$ surfaces shown in Fig. 13 indicate that there was very strong vertical wind shear, particularly below 70 kPa. The shear supported the intense turbulence in the frontal zone that was experienced during the low-level aircraft flight legs. The $\theta_e$ field indicates a horizontal gradient of about 6°C (100 km)~1 in the boundary-layer air. A large portion of the atmosphere was close to neutral stability for CSI. The combination of strong vertical shear and very weak vertical gradients of $\theta_e$ in the region 0–80 km between 80 and 70 kPa resulted in instability for CSI. Unstable conditions were also indicated in the cold air between 75 and 150 km at 65 kPa. The importance of strong wind shear on the stability properties is shown in Fig. 14. While the vertical sounding is clearly stable for cumulus convection, the slanted sounding is unstable for CSI, even when allowance is made for the drag effect of the condensed water.

The infrared GOES satellite picture imagery for 1801 UTC 24 February 1989 displayed in Hartnett and Browne (1989, p. 201) shows that the analyzed section was covered with high clouds. There was a general alignment of cloud patterns in the north–south direction, that is, in the shear direction; however, the satellite imagery resolution is too coarse to allow for a detailed comparison with our stability analysis. No radar data were sampled around 1800 UTC.

Another opportunity for analysis occurred at 2100 UTC. This cross section (Fig. 15) was sampled in the cool air about 300 km northwest of the surface low center. The atmosphere was found stable for slantwise convection below 80 kPa ($q_v > 0$). Between 80 and 70 kPa, patches of slantwise neutrality occurred. A sounding analysis (not shown) confirms that the air along $M$ surfaces was stable, yet less stable than the associated vertical soundings. The results suggests that the air northwest of the surface low is stable for CSI, despite the strong vertical shear.

9. Conclusions and discussion

During the ERICA field study, LeSondes were released from research aircraft to sample the temperature, humidity, and velocity in the vicinity of cyclones over the ocean. Six cases during four different intensive observing periods were selected to construct vertical sections perpendicular to the low-level wind shear. To assess the stability properties for CSI, a sounding anal-

![Fig. 12. Skew $T$-log $p$ diagrams showing $T_v$ and $T_w$ along selected paths at about 0200 UTC 13 February 1989. (a) Slanted sounding along $M = 0 \text{ m s}^{-1}$ surface. (b) Vertical sounding passing through $M = 0 \text{ m s}^{-1}$ at LCL. (c) Slanted sounding along $M = 20 \text{ m s}^{-1}$ surface. (d) Vertical sounding passing through $M = 20 \text{ m s}^{-1}$ at LCL.](image-url)
ysis was carried out along surfaces of constant $M$. The buoyancy of the parcel relative to the environmental air was computed both with and without the effect of the weight of the condensed water on the density. Liquid water content (LWC) measurements indicate that a LWC > 0.5 g m$^{-3}$ was not uncommon for ERICA storms. In addition to the loading of the small cloud water droplets, the loading of the rain and ice particles has to be considered. For such conditions, it seems reasonable to approximate the total water content using its adiabatic value (Xu and Emanuel 1989). The virtual temperatures with and without the water loading effects differed by about 2°C at midlevels, which is a typical value of observed buoyancy in active convective clouds.

The stability analysis based on $(\theta_v, M)$ and $\varphi_v$ sections neglects the water loading effect and thus has a strong bias toward overestimating the amount of instability for slantwise convection. Thus, we strongly support Xu and Emanuel (1989) in their exhortation to include the effect of the condensed water when analyzing parcel instability. For the stability analysis, we have usually chosen a parcel originating from near the surface. Parcels starting from higher levels usually would have been less unstable, but more slantwise instability is not ruled out either. The major findings of our analysis are the following.

1) The warm air of the warm-frontal zone ahead of the developing IOP 4 cyclone is conditionally stable or neutral with respect to cumulus convection when the water drag is included; however, the region is unstable for slantwise convection throughout the 90–50-kPa layer.

2) Twelve hours prior to rapid deepening of the IOP 2 cyclone, the region ahead of the cyclone track had a boundary layer (from surface to about 80 kPa) that was neutral or slightly stable for upright convection. However, this boundary layer was unstable for slantwise convection. The layer with strong shear between about 80–70 kPa had even larger upward buoyancy values along constant-$M$ surfaces, despite the fact that this layer was clearly stable for vertical convection.

3) At the stage of explosive cyclogenesis of both IOP 2 and the IOP 4, deep plumes of convectively unstable air were found in the warm sector close to the surface.
low center. In both these “bombs,” the convective instabilities were caused and apparently sustained for some time by the high-$\theta_e$ air found in the marine boundary layer.

4) The results for IOP 8 indicate that slantwise convection throughout a deep layer (that is stable for up-right convection) can occur in a developing cyclone that fails to show explosive growth.

5) The frontal zone during IOP 7 was associated with very strong vertical shear and thereby provides favorable conditions for the occurrence of slantwise instability. Our analysis confirms that the air was unstable for CSI yet stable for cumulus convection. Heavy precipitation was indicated on the radar.

6) Stable conditions for slantwise convection are indicated in the cold sector of the cyclones that we have analyzed.

These findings need to be confirmed by examining more marine storm cases. Sounding data with even higher spatial resolution than available in ERICA may be required to further reduce the uncertainty introduced by the horizontal interpolation between adjacent soundings. However, we feel rather confident that further studies will provide even more evidence on the existence of CSI in the warm frontal zones of developing marine cyclones. Furthermore, we speculate that these regions of CSI have an important impact on the development of marine cyclones by stimulating the interaction of low-level and upper-level baroclinic processes. Before being more specific on the details of our speculation, some results of previous studies are briefly outlined.

First, we note that even if the atmosphere is stable with respect to CSI, the degree of stability (i.e., the magnitude of equivalent potential vorticity) can determine the rate of frontogenesis resulting from a given amount of forcing. The role of potential vorticity in baroclinic processes has been studied in the framework of semigeostrophic theory (Hoskins and Bretherton 1972). In semigeostrophic frontogenesis (Hoskins 1982) slanted updrafts are forced whenever there is positive advection of semigeostrophic vorticity by the thermal wind. Furthermore, the vertical motion is inversely proportional to the potential vorticity because potential vorticity acts like stability in the diagnostic Sawyer–Eliassen equation for the ageostrophic circulation. The effects of moisture (during reversible thermodynamic processes) can be included by replacing potential vorticity in the analysis with $q_e$. Emanuel (1985) has argued that a dramatic interaction between midlevel and low-level processes will occur when $q_e$ approaches zero in a saturated frontal zone. Numerical simulation results (Thorpe and Emanuel 1985) have indicated that the rate of surface frontogenesis is indeed sharply increased by the release of latent heat from slantwise convection.

Since frontogenesis is often present during cyclogenesis, the degree of slantwise stability is also a crucial factor in the development of extratropical cyclones. Observations suggest that cyclogenesis in a moist, slantwise unstable environment is very intense (Reed and Albright 1986). Numerical simulations of explosive cyclogenesis support the notion that slantwise convection strengthens warm frontal development and contributes to the intensification of surface vorticity (Nordeng 1987; Kuo and Reed 1988; Mailhot and Chouinard 1989; Kuo and Low-Nam 1990; and Hedley and Yau 1991). Indeed, our findings for IOP 2 and IOP 4 are consistent with Fig. 10 in Hedley and Yau (1991), which shows a convectively unstable region caused by surface fluxes south of the surface warm front. This instability is released in slantwise ascent along the deep warm-frontal zone marked by slantwise neutrality or instability.

With this background, we speculate that the slantwise ascending motion associated with CSI can lead to a rapid spinup of low-level vorticity because it results in rapid frontogenesis and vortex-tube stretching and thereby fills the storm center with air of high vorticity. Future observational programs and numerical modeling studies will indicate to what extent this suggested spinup process occurs in marine cyclones.

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