Influence of the Coriolis Force on Two-Dimensional Model Storms

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(Manuscript received 3 January 1990, in final form 16 August 1990)

ABSTRACT

Two-dimensional model simulations were made to gauge the effect of the Coriolis force on model squall lines. The case chosen for intensive study had low-to-moderate wind shear confined to low levels. With this wind shear, two Coriolis simulations were made, with and without a geostrophically balanced along-line temperature gradient. Additional simulations were made with other wind shear intensities to test the sensitivity to low-level shear.

Unlike their nonrotational counterparts, none of the Coriolis model storms were able to attain or maintain a "quasi-equilibrium" state. Quasi-equilibrium storms possess mature phases characterized by essentially statistically steady behavior with respect to storm strength, propagation speed, etc. Instead, the Coriolis storms possessed mature phases marked by gradual but definite decay. These are the first model storms created with the present model sounding and wind profiles that have terminal mature phases due to physically realistic forcings. However, the time scale of the decay, at least in these cases, makes it unlikely that Coriolis forcing is the primary mechanism behind the demise of real long-lived, mature squall line thunderstorms.

In each rotational case, the decay phase was marked by two major temporal trends absent in the mature phase of the nonrotational simulations: the continued contamination of the forward environment with storm-induced subsidence warming and the decline in intensity of the rear inflow current. The subsidence warming was slowly eradicating the convective instability of the air flowing into the storm, and the dissipating inflow current appeared to be at least partially responsible for the progressive collapse of the storm's subcloud cold pool. The accumulation of subsidence warming was clearly injuring the model storm. The role that the declining rear inflow played in the decay phase is less clear and requires additional study.

It was found that the inclusion of the geostrophically balanced along-line temperature gradient had small but measurable consequences in this situation. Warm advection at low levels ahead of the storm worked to negate the effect of warm advection aloft on the convective instability, and cold advection into the cold pool opposed the general decline in pool intensity. The net effect was that the Coriolis-associated mature-phase decaying tendency was slowed somewhat, but not arrested.

1. Introduction

In two recent papers, Fovell and Ogura (1988, 1989) described their attempts to replicate the basic temporal behavior and spatial structure of a typical midlatitude squall line using a two-dimensional (2D) model. In the former (hereafter FO88), the employment of initial vertical thermodynamic and wind profiles similar to those existing prior to the onset of convection in Oklahoma on 22 May 1976 resulted in the establishment of a model storm that, after a period of organization, settled down into a realistic-looking mature state that appeared to have no end; this was referred to as the "quasi-equilibrium" state. Quasi-equilibrium storms are characterized by essentially statistically steady behavior with respect to storm strength, propagation speed, and other characteristics when considered over periods longer than those of the individual cells during the mature phase. In the latter paper (hereafter FO89), the same thermodynamic conditions were used but the low-level wind shear was varied over a wide range of intensities. Most of those storms also achieved the essentially periodic and apparently endless mature phase.

A number of factors, numerical and physical, was identified as responsible for the development and maintenance of the statistically stable mature phase. Numerically, a very wide computational domain was used; a central, high-resolution area was flanked on each side by a stretched grid zone that served to both contain the storm and push the lateral boundaries away from the storm. The problems encountered when small domains were used were discussed in both papers. Dynamically, very favorable, horizontally homogeneous and temporally invariant environmental conditions were employed. Thus, potentially important complicating factors such as large- and small-scale variations in environmental conditions and diurnal forcings have been ignored. Two-dimensionality implies that the possibly important third dimension was also neglected. In addition, the very long lifetimes (and model integrations) of the Fovell and Ogura storms raise the pos-
sibility that rotational influences due to the Coriolis force could be important.

Summarizing the results of both papers, Fovell and Ogura (1989) came to the conclusion that, given the limited model framework and favorable and persistent environmental conditions employed, the seeds of the storm system’s demise were not contained within the storm itself. The storms were perhaps not as strong as they could have been—that is, they were in a “suboptimal” condition, using Rotunno et al.’s (1988) terminology—and yet were able to persist in this theoretically less favorable state indefinitely. On the basis of their sensitivity tests, FO88 and FO89 remarked that there may be numerical reasons why model storms created using similar initial conditions in other 2D models (especially those utilizing very limited domains) had finite life spans.

Storms in reality die. The research plan has been to relax one model restriction at a time, to see if we can isolate the factor(s) necessary to bring our well-organized, intense storms to an end. As a first step in that direction, we rewrote the model to include the alongline wind component that could be induced by the Coriolis force. The model is still in no sense three-dimensional since pressure variations are not permitted in the along-line direction. Chosen for intensive study was a wind shear case considered by FO89, possessing 7.5 m s⁻¹ of wind speed change over the lowest 2.5 km, and a value of the Coriolis parameter corresponding to 40°N.

The importance of the Coriolis force is that it imposes upon the model storm a fundamental length scale, the Rossby radius of deformation (cf. Gill 1982), which did not exist before. The rotation of the earth acts on gravity waves excited by the convection by controlling their ability to propagate away, effectively trapping the compensating subsiding motion they carry. This not only forces the concentration of warming associated with the subsidence within the storm’s locality, but also causes the warming to accumulate with time. The former effect implies that the feedback between the storm and its environment could become more important in determining how the storm organizes; the latter effect should work against the storm’s health, probably preventing it from achieving the same, nicely periodic mature phase as the nonrotational case, possibly resulting in the storm’s demise. This trapping and concentration of the subsidence warming in the cloud’s locality is predicted by the linear, nonprecipitating model of Bretherton (1987).

The Coriolis simulations described herein have also answered and raised some questions about the “rear inflow” current (cf. Smull and Houze 1987). This current is an interesting and controversial phenomenon, sometimes blamed for injuring the storm and other times credited with aiding it. For example, Smull and Houze (1985) noted that the appearance of the intense current at the rear of the 10–11 June 1985 squall system coincided with the rapid decay of much of its convective region. This observation has sometimes been taken to imply a causal relationship between the current intensification and storm dissipation. However, Smull and Houze reported further that the portion of the line that most likely was directly affected by the rear inflow intensified and surged ahead from the rest of the line. This observation would support the notion that a strong rear inflow current is beneficial to the convective portion of the storm. To complicate things further, Zhang and Gao (1989) speculated that the rear inflow may only aid the storm while the environment is favorable and injure the storm when those conditions worsen.

2. Model description and initial conditions

The model used in FO88, FO89 and herein is a 2D model built using the anelastic equation set. The leapfrog scheme with a time step of 5 s is used to compute advection, and a block tridiagonal solver is used to invert the diagnostic pressure equation. Stretched grid meshes are used in both directions, in the vertical to enhance resolution at low levels where it is most required (minimum grid spacing 200 m) and in the horizontal to push away the lateral boundaries from the region of interest. The model grid is 40 points deep (model top 22 km) between rigid plates and 595 points wide (physical width 4778 km). The central 315 grid points in the horizontal represent the high-resolution (grid spacing 1 km) zone where the most intense portions of the storm and its environmental response are intended to reside; outside of this zone, the grid is stretched at a rate of 1.0305:1. This stretching rate is equal to the gentlest used thus far in the model.

The model was initialized with the same temperature and moisture profiles used by FO88 and FO89. These profiles were adapted from environmental conditions existing prior to the onset of convection in Oklahoma on 22 May 1976 and are generally similar to conditions typical of severe springtime convection in Oklahoma (cf. Bluestein and Jain 1985). As in FO89, we have presently considered only cases with linear wind shear between the surface and 2.5 km above ground level. The case chosen for intensive study had 7.5 m s⁻¹ of wind speed change over that height (i.e., Δu = 7.5), for a shear intensity of 3 × 10⁻³ s⁻¹. Within the context of the FO89 study, it is appropriate to characterize this case as having wind shear of low-to-moderate intensity. It was selected because its nonrotational counterpart displayed one of the most perfect examples of quasi-equilibrium behavior the model has produced. It was found in that work that simulations with smaller low-level wind shear intensities tended to be less organized and more aperiodic and those with higher shears tended to evince more complex (yet still essentially periodic) behavior. One of FO89’s moderate-to-high shear cases, with 20 m s⁻¹ of low-level wind speed change (Δu = 20), will also be examined for purposes of comparison with the Δu = 7.5 case.
3. Results

a. Experimental strategy

The addition of the Coriolis force to the numerical model entails the modification of the two horizontal momentum equations and also the thermodynamic equation. The terms added to the momentum equations are \( +f_0u \) and \( -f_0u' \) to the westerly, across-line \( (u) \), and the northerly, along-line \( (v) \) component equations, respectively. The variable \( u' \) refers to perturbations from the undisturbed initial state, which varies with height \( [i.e., u'(x,z) = u(x,z) - u_0(z)] \), and \( f \) is the Coriolis parameter \( (\sim 10^{-4} \text{ at } 40^\circ \text{N}) \). The height dependence of the across-line wind geostrophically implies the existence of a virtual along-line temperature gradient that could be advected by the along-line wind. This forcing is represented as \(-v f_0 \theta_m(z) g^{-1}(\partial u_m/\partial z)\) by utilizing the thermal wind relation.

Two simulations using the \( \Delta u = 7.5 \text{ wind profile were made with the Coriolis force activated and will be described in detail below. The first, discussed in subsections } b \text{ and } c \text{, ignored the thermodynamic geostrophic balance term as a first approximation and will be referred to as the “unbalanced” run. Neglecting the geostrophic thermodynamic term has allowed us to not only assess the impact of activating the Coriolis force due to momentum forcing alone but also gage the relative importance of the geostrophic shear balance term on the model storm. The specific import of including the along-line temperature gradient is the topic of subsection } d \text{. In addition, some simulations have been made in an attempt to gage the effect of varying the low-level shear intensity on the results; in particular, the } \Delta u = 20 \text{ case will be discussed in subsection } e \text{.}

b. Time history of the unbalanced Coriolis simulation and comparison to control

The unbalanced model with low-to-moderate low-level shear was integrated for 17 h (1020 min), which is approximately equal to one inertial cycle \( (i.e., 2\pi f^{-1}) \). Figure 1a shows the time history of the domain maximum vertical velocity for this simulation. For comparison, the time history of the control \( (\text{nonrotational}) \) run is also presented (Fig. 1b). (The control run differs from the simulation of FO89 in that the present run utilized the numerically wider horizontal grid of this study and the simulation was extended.) Very early in the simulation the two histories were quite similar. Both storms had created well-developed subcloud cold pools and rising, upshear-oriented airflows typical of multicellular squall line storms by about 200 min. Establishment of these features is considered the beginning of the mature phase that, in the control run, was characterized by a quite stable, extremely repetitive oscillation in the value of the maximum updraft as well as in other fields. Each updraft peak during the mature phase was associated with a different cell, which was created within the rising front-to-rear airflow over the forward edge of the surface cold pool \( (\text{the “gust front”}) \) and subsequently was swept towards the rear, to be soon replaced by yet another cell. The period of the mature phase oscillation was about 16 min and continued for the duration of the simulation. As noted in FO89, the small decay tendency in the storm, detectable in Fig. 1b at large times among the maxima of successive cells, was in fact spurious as it was lessened, as still more gently stretched grid designs were adopted.

It is immediately apparent that this Coriolis model storm attained a mature phase that was weaker, more chaotic, and less stable than that achieved by the control. In addition, unlike the control storm, the Coriolis storm was unable to maintain the original temporal characteristics it developed early in its mature phase. Instead, there was a dramatic transition to a still weaker, more rapidly oscillating state after about 550 min. This immature phase transition was soon followed by a sudden slowing in the propagation speed of the storm, from the 13.4 \( \text{m s}^{-1} \) value it was before 570 min to about 12.2 \( \text{m s}^{-1} \) shortly thereafter. For comparison, the control model storm attained a propagation speed of 14.8 \( \text{m s}^{-1} \) early in its mature phase, which remained constant after that time.

The propagation speed of the post-transition model storm continued to decrease gradually for the remainder of the simulation, dropping by another 1.2 \( \text{m s}^{-1} \) by 1020 min. This slowing occurred while the storm itself was declining in intensity at a gradual yet significant rate. The weakening is better seen in the time histories of condensation and rainfall \( (\text{Figs. 2a,b, respectively}) \) than in the maximum vertical velocity, which depicts the storm state at only one grid point for each time plotted. The Coriolis storm’s behavior stands in sharp contrast to that of the control, which also experienced a \( (\text{much smaller}) \) decline in strength \( (\text{due to numerical rather than physical reasons}) \). The decaying trend in the Coriolis simulation survived sensitivity testing of the fashion described in FO88 and FO89 and therefore, even though it is quite gradual, this storm represents the first model storm created from the present model sounding that decays for physically realistic rather than numerical reasons. Some of the results of these tests will be discussed in later sections, where appropriate. The mechanisms behind this decline are interesting, but it appears certain that due to the extensive time scale needed to injure the storm, we have not found the primary force that brings about the demise of mature storms in the real world.

c. Storm structure through the mature phase in the unbalanced Coriolis simulation

To more closely examine the effect of the Coriolis force in general, and the mature phase transition in particular, the following figures present selected model
FIG. 1. Time histories of domain maximum vertical velocity each time step for the 17-h integrations using the low-to-moderate shear ($\Delta u = 7.5$) wind profile for (a) the unbalanced Coriolis run and (b) the nonrotational control run.
fields, averaged over time in reference frames fixed with the leading edge of the storm in the manner previously adopted in FO88 and FO89, for three periods during the unbalanced Coriolis storm’s mature phase. The averaging intervals chosen were 270–330 min (prior to the transition), 570–630 min (shortly after the transition), and 870–930 (long after the transition). In all of these fields, the dark, solid line represents the outline of the subcloud cold pool, separating the evaporatively chilled air near the surface from the condensationally warmed air in the cloud above. Only the tropospheric portion of the domain is shown. Some of the figures present the entire (untransformed) domain while others zoom in on a portion of the fine grid zone to focus on the convective region.

1) THERMODYNAMIC AND ACROSS-LINE WIND FIELDS

Figure 3 presents the potential temperature fields, expressed as deviations from the initial state, for the three averaging intervals. Time-averaged fields for the control run, as they existed at 5 and 15 h, are presented for comparison purposes in Fig. 4. The temperature perturbation distribution of the early mature phase period of the Coriolis simulation (Fig. 3a) looks quite similar to those previously produced by the Fovell–Ogura model as well as other cloud models in general, and that of the control run in particular. Present was a convective region consisting of large perturbations, with condensational warming overlying a deep pool of evaporationally cooled air, residing ahead of a trailing region marked by a deep layer of subsidence-associated warming and a shallow surface cold pool. Subsidence warming also extended for some distance ahead of the leading edge of the storm, confined to high levels for the control simulation.

We have already seen that a striking difference between the Coriolis and control simulations was that the former was unable to attain statistical steadiness during its mature phase like the latter. Although separated by 10 h, the fields for the control run are remarkably similar, especially in the central fine-grid region, another demonstration of its steadiness (compare Figs. 4a,b). The storm’s trailing subsidence region did expand rearward somewhat, as did its cloud shield (not shown), but the major features of the model storm have not significantly changed in intensity during the passage of a very long period of time. In contrast, as time progressed in the Coriolis simulation (Figs. 3b,c) both the size and the intensity of the subsidence zones increased while the cold pool itself shrank. By late in the simulation, the accumulation of heating in the domain had become quite significant. It will be demonstrated later that on the forward side of the storm, the accumulating midlevel subsidence warming was gradually eroding the convective instability of the environment as measured by its CAPE (convective available potential energy) with respect to a near-surface air parcel.

Expansion of the flanking subsidence zones in the Coriolis model storm occurred as the storm itself was slowing and its circulation became weaker. Figure 5 shows fields of the storm-relative across-line horizontal wind for the three periods. For these panels, we focus more closely on the convective zone. As the cold pool declined in depth, the storm-relative front-to-rear airflow over the pool became weaker. At the same time, the intensity of the rear inflow current that underlay the front-to-rear jet decreased in strength and extent.

The startup conditions employed mean that the model storm’s rear inflow current is produced solely by the storm itself. From the back of the storm, the typical air parcel in the rear inflow traverses the trailing subsidence region (above the pool’s shallow “body”)

![Net Condensation by hour](image1)

![Total Precipitation by hour](image2)

**Fig. 2.** Bar graphs showing production of (a) net condensation and (b) total rainfall each hour by the low-to-moderate shear nonrotational control (striped bars) and unbalanced Coriolis (solid bars) model storms. (Units are $10^{-5}$ kg m$^{-1}$ h$^{-1}$.)
and eventually descends into the pool in its deep "head" zone (see FO88; FO89's Fig. 27). The direct effect of the current, then, is to import the warm and dry air of the trailing region into the cold pool. Once inside the pool, the parcel is chilled and moistened by rainwater evaporation and must eventually be evacuated from the head zone to make way for additional inflowing air. This is most efficiently accomplished by the surface return flow that exhausts air rearward into the cold pool's trailing body. Especially in low shear cases, however, part of the evacuation is accomplished by mixing the pool air out through the pool top into the cloud's front-to-rear flow (e.g., FO89, Fig. 22). This may well injure the storm's updraft, but its contribution diminishes as the rear inflow elevates and thus more of the vented air is borne by the along-surface component. The ultimate source of much of this inflow current's air is the front-to-rear flow itself (e.g., FO88, Fig. 11f). These processes, as well as all others operating within the storm, must be in balance on the quasi-equilibrium time scale for statistical steadiness (whether at a relatively strong or weak state) to occur.

Trailing region air is both warm and dry but its dryness appears to be the more important, if only because the horizontal gradients in that field leading into the cold pool are substantially greater. Figure 6 shows fields of perturbation vapor mixing ratio that existed during the three mature phase periods for the same reference.
frame as Fig. 5. As is both typical and logical, drying existed at low levels throughout the subsidence region and was maximized just behind the cold pool head. It is there that the still relatively intense cell downdrafts had been sufficiently depleted of precipitation particles that evaporational cooling of those particles could no longer dominate the warming and drying due to subsidence. This dry air was being imported into the cold pool by the rear inflow current. The panels of the figure show that although the magnitude of the drying was increasing with time, the actual flux of dry air into the pool in the current was actually declining since the intensity of the inflow current was decreasing. In fact, by the late mature phase period (Fig. 5c) there was virtually no rear-to-front flow relative to the storm at low levels outside of the cold pool and so the importation of dry air from the rear was very small.

There is, of course, another way for dry air to reach the cold pool; that is, from the forward side in discrete bursts between the cells (see section 4a). This requires unsteadiness on the part of the storm but can account for an appreciable amount of the cold pool’s dry-air intake in low-to-moderate shear storms that exhibit considerable unsteadiness on the cell-lifetime time scale (see FO88). We can be sure, however, that the net dry advection into the cold pool in this Coriolis storm was decreasing with time during the mature phase. The major indicator and result of this decline was the progressive moistening of the cold-pool air. Though the dry advection was decreasing with time, the evaporation of rainwater within the pool was continuing, apace with the result that the pool air was gradually moistening. Figure 6 shows that between the early and late mature phase periods, there has been a considerable relative as well as absolute moistening of the subcloud air. Since the humidity of the pool air was increasing, it is unsurprising that the evaporation of rainwater in the pool was declining, since the evaporation rate is proportional to the air’s dryness. The slow decline in evaporational cooling produced in the pool led to the slowing of the model storm after the mature phase transition since it is the weight (i.e., the buoyancy deficit) of the air that was hydrostatically driving the storm’s cold pool (and thus the storm itself) forward.

2) The Coriolis-induced along-line wind field

The most direct effect of the inclusion of the Coriolis force is to introduce an along-line wind component created by the across-line wind, which is attempting to achieve a balance with the horizontal pressure field. Figure 7 shows the distribution of the along-line wind
for the mature phase-averaging intervals. Since the Coriolis parameter is taken to be constant across the simulation domain, it follows that the along-line direction must be north–south. Also, one should remember that the force is Galilean invariant in the model framework, and thus specifically acts on perturbations of ground-relative (rather than storm-relative) winds. In order to visualize ground-relative airflow in Fig. 5, 13.4, 12.2, and 11 m s$^{-1}$ must be added to panels a, b, and c, respectively. Perturbations are then found by subtracting the base state wind profile.

As expected, the effect of the ground-relative (and storm-relative) easterly perturbations in the storm’s front-to-rear airflow is to induce a southerly wind that has served to deplete the easterly airflow of momentum, causing it to decelerate. Likewise, on the forward side, the westerly winds in the high-level forward anvil outflow and low-level rear inflow (which are fairly large perturbations when considered in a ground-relative reference frame) have developed northerly wind components associated with them. Note the rearward translation with time of the couplet of upper-tropospheric southerlies and near-surface northerlies, located in the zone above $x = -125$ early in the mature phase (Fig. 7a). The distortion evident in the couplet with time is due to its propagation into the improperly mapped stretched grid zone ($x < -150$ km). The strength of the near-surface element of the couplet de-
increases since it is being opposed by the Coriolis acceleration of the across-line ground-relative wind perturbations (which are easterlies there). In addition, it is being damped by physical diffusion and also the implicit numerical diffusion associated with entering into and traveling through the expanding mesh. On the other hand, the intensity of the upper-tropospheric southerlies has remained approximately constant over time; the Coriolis acceleration of the general easterly perturbations are helping the feature to resist diffusion.

The propagation of this couplet away from the leading edge of the storm gives one the picture that the storm was still continuing to extend its influence slowly towards the rear, even though it was weakening with time. There is also a suggestion of forward expansion in the panels of Figs. 5 and 6 as well. Note that the reach of the storm-relative westerly forward anvil outflow increased with time, inducing more intense and more extensive northerlies. Indeed, the intensification of the northerlies within the forward anvil in the immediate vicinity of the storm is quite interesting. The acceleration of the northerlies there (about 3 m s⁻¹ per 300 min) is roughly as expected from the Coriolis acceleration of persistent ground-relative perturbations, which averaged about 3–4 m s⁻¹ in the zone early in the mature phase, after allowing for the effects of advection and diffusion that were in opposition there. These opposing effects were even stronger in the cold-pool portion of the rear inflow current and prevented any growth in the northerly wind component there.
3) THE PRESSURE FIELD

Three elements of the surface pressure distribution typically associated with squall lines are the wake low (e.g., Johnson and Hamilton 1988), the mesohigh (e.g., Fujita 1955), and the presquall low (e.g., Hoxit et al. 1976). The presence of these features, and their evolution through the mature phase, can be seen in Fig. 8. The mesohigh is typically located in the heavy precipitation zone following the passage of the gust front and is hydrostatically produced by the weight of the evaporatively chilled air in the cold pool bearing down on the surface. FO88 divided the pressure field into its buoyant and dynamic components and found that the total pressure in the surface mesohigh is smaller than that due to the buoyancy-forced component alone, since the dynamic portion of the pressure was in opposition to it. The dynamic component was associated with the acceleration of air in the descending rear inflow current as the current entered the cold pool and approached the surface (see, for example, FO88’s Figs. 11c,e, and 12c,d).

Unlike the control simulation in particular and the model storms described in FO88 and FO89 in general, the surface mesohigh in the Coriolis storm was not composed of absolutely positive values of perturbation pressure during most of the mature phase, although clearly, the high was a local maximum of pressure. As discussed in FO88, the pressure field is obtained in the anelastic model from the inversion of a diagnostic el-
Fig. 8. Time-averaged fields of perturbation pressure from the low-to-moderate shear unbalanced Coriolis model storm for the same averaging intervals as Fig. 3. Perturbations are relative to the initial, horizontally homogeneous state; a cautionary note is discussed in the text. Contour interval is 0.4 mb. As in Fig. 3, the entire horizontal width of the model troposphere is presented and the stretched grid resides at $|x| > 150$ km. Again, the outline of the cold pool has been drawn on each panel.

Liptich equation that, due to the boundary conditions employed, yields perturbation estimates only to within an unknown constant. The constant could be different for each time step for physical as well as numerical reasons, but the value of the constant itself is irrelevant to the model since only spatial derivatives of the pressure field are needed. The existence of a spurious numerical trend could be one reason why the value of the highest surface perturbation pressure became more and more negative with time; although the fact that the domain was definitely heating up at midlevels had to be the most important factor. Also, as the cold pool degraded in depth and intensity, there was progressively less cold air to hydrostatically change the low pressure perched at the top of the cold pool into an absolutely high value at the surface.

The other two features, the wake and presquall lows, intensified with time, again in a manner that was hydrostatically consistent with the accumulation of warming in the model domain. Observations have led us to expect this (e.g., Johnson and Hamilton 1988). The presquall low is the general region of locally minimum surface pressure ahead of the storm at about $x = +75$ km in the panels of Fig. 8, while the wake low is detectable behind the system at around $x = -125$ km—especially in the latter two panels. The dramatic rise in intensity of the wake low was due in part to the shrinkage of the cold pool. Early in the simulation, the
horizontally extensive cold pool kept the hydrostatically
determined low pressure zone in the trailing region off
the ground. As the simulation progressed, the tail of
the cold pool retreated back towards the convective
region, allowing the low pressure zone to descend to
the surface, creating the wake low. This is one of the
mechanisms that Johnson et al. (1989) described.
Again, it is the Coriolis storm’s inability to attain sta-
tistical steadiness that has brought this about.

d. Import of the geostrophic balance term

Next, the geostrophic balance term was activated in
the thermodynamic equation. Generally, the balanced
Coriolis model storm was found to be quite similar to
the unbalanced simulation already presented. The
storm still decayed in strength as it slowed in speed,
although both rates were more gradual than in the un-
balanced simulation. The storm also had a marked,
abrupt less dramatic, mature phase transition. Time-
averaged model fields for this case were sufficiently
similar to those already shown so that they need not
be presented.

The advection of the geostrophically balanced along-
line temperature gradient attempted to oppose or mit-
igate the mature phase tendencies concerning the in-
stability of the forward environment and the integrity
of the cold pool that were noted in the unbalanced run.
Figure 9 presents the average advection rate being ex-
perienced by the storm during a time interval well in
the mature phase (870–930 min). The influence of the
thermodynamic term is confined to the lowest 2.5 km,
since that is where the shear was concentrated in this
case.

On the rear side of the storm, the rear inflow current
still slackened in intensity with time but the weakening
of the subcloud pool was partially offset by cold ad-
vection associated with the induced northerly winds
within the pool. Consequently, the balanced storm
propagated somewhat faster than the unbalanced storm
during its mature phase (1.3 m s\(^{-1}\) faster at 900 min)
and slowed at a more gentle rate with time. On the
forward side, warming induced at low levels by south-
eroies was attempting to mitigate the stabilizing effect
of subsidence warming aloft. As a result, the decline
in CAPE with time was slightly smaller in this run. (In
fact, the low-level warm advection was not maximized
right at the ground so there was also a detectable, but
unimportant, increase in the convective inhibition in
the forward environment over that in the other Corio-
lis run).

The net effect of the addition of the balance term
was to make the balanced Coriolis model storm mar-
ginally stronger and its decaying tendency slightly
slower compared to the unbalanced storm. Figure 10
compares condensation and rainfall production be-
 tween the two Coriolis storms. Indeed, around 17 h it
appeared that the balanced model storm was trying to
rebound from its decline, which is why this integration
was extended for another three hours. It appears, how-
ever, that the departure from the general downward
trend was temporary. In any case, the further accu-
mulation of destructive subsidence warming with time
would seem to preclude any hope of eventually attain-
ing a stabilized state like the control run.

Therefore, the forces directly or indirectly associ-
ated with the balance term were insufficiently strong in
this case to overcome the deleterious consequences that
adding rotation to the model framework produced.
However, we note that it is possible to find situations
in which that term would be of greater relative impor-
tance, in both the forward environment and the sub-
cloud pool. Also, it is equally possible to design shear
profiles that would cause the geostrophic advection in

\[ \Delta u = 7.5 \]

![Coriolis (balanced)](image)

**Fig. 9.** Average advection rate for the low-to-moderate shear Coriolis model storm with the geostrophic thermal wind balance term in the thermodynamic equation activated. Contour interval is 0.06 K h\(^{-1}\). Averaging interval was 870–930 min, corresponding to the last of the three intervals considered in the analysis of the unbalanced case. A 90-km portion of the central fine grid is displayed and the outline of the storm’s cold pool has been drawn on the panel; labels on the x axis are in kilometers (km) from the domain center. Due to differing propagation and domain translation speeds for this case, the model storm resides at a slightly different position relative to the domain center than the unbalanced run did.
Like the nonrotational $\Delta u = 7.5$ run, the control model storm attained a mature phase (after 4 h) characterized by both persistence and approximate statistical steadiness. FO89 discussed that this storm produced individual cells in sets with a basic repeat cycle of about 42 min, with smaller amplitude and longer period oscillations superimposed. In contrast, the Coriolis model storm, which started out as strong as the control run, was found to progressively weaken. In agreement with the results for the lower shear case, the rotational storm propagated more slowly than the control run and also slowed with time. While the control run achieved a steady speed of 22.8 m s$^{-1}$ for the entire (unending) length of its mature phase, the Coriolis model storm's speed went from 22.0 m s$^{-1}$ at 300 min, down to 21.4 m s$^{-1}$ by 600 min, and down further to 19.8 m s$^{-1}$ by 900 min.

Also generally similar between the lower and higher shear simulations was the impact of rotation on the potential temperature field, although there are some interesting differences. Figure 12 shows time-averaged data for this field for a time interval ending at 900 min for the Coriolis [panel (a)] and control [panel (b)] storms. Compared to the control run, the Coriolis storm had both a weaker, shallower cold pool and stronger, more concentrated warming at midlevels, characteristics also found in the lower shear simulation set. However, note that the rotational storm's accumulation of subsidence warming on its forward side was substantially larger than that found in the larger shear control run and both of the lower shear runs. In this storm, the influence of the Coriolis force was able to reach a greater distance upstream. The time-averaged fields of storm-relative across-line ($u$) and along-line ($v$) velocity fields (Figs. 13a,b) for this Coriolis case also look familiar, but the latter especially emphasizes

development of the low-level temperature inversion.

e. Sensitivity to magnitude of low-level wind shear

A sample of wind shear intensities were chosen to assess sensitivity to the magnitude of the low-level shear. Below, we compare control and unbalanced rotational simulations for the $\Delta u = 20$ profile, a moderate-to-high shear case. The simulations were run for 15 h (900 min).

In most respects, the response of the more highly sheared storm to activation of the Coriolis force was qualitatively similar to that for the lower shear situation. Figure 11 presents a time history of net condensation for the nonrotational and Coriolis simulations.

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**Fig. 10.** Line graphs showing production of (a) net condensation and (b) total rainfall each hour by the balanced (indicated by diamonds) and unbalanced Coriolis (indicated by squares) model storms with low-to-moderate wind shear. The information for the latter simulation was presented in a different fashion in Fig. 2. The balanced model storm was carried out 3 additional hours, to 20 h. (Units are $10^7$ kg m$^{-1}$ s$^{-1}$.)

**Fig. 11.** Bar graph showing production of net condensation each hour by the moderate-to-high shear ($\Delta u = 20$) nonrotational control (striped bars) and unbalanced Coriolis (solid bars) model storms. (Units are $10^7$ kg m$^{-1}$ s$^{-1}$.)
that the Coriolis-related atmospheric adjustment was much less biased towards the rear of the storm. (Compare the positions of the northerlies in the forward anvil region of Figs. 6 and 13b.)

The disparity in storm strength between the control and Coriolis model storms was smaller in the higher shear situation and its decay rate was also less pronounced, despite the fact that its along-line circulation was considerably stronger (compare Fig. 11 with Fig. 2a) and the subsidence-induced stabilization in its forward environment was greater. Since the higher shear storm was stronger to start with, it apparently was even more resistant to decay (or perhaps just change in general, for better or worse) within the time span of the simulation. For example, its cold pool was stronger and deeper to start with, and was thus less sensitive to the deleterious effects of rotation. While the cold pool of the lower shear case was progressively being starved of dry air due to the decline of its rear inflow, the higher shear storm’s pool was still being fed with comparatively drier (see Fig. 13c) trailing region air at a considerable (although diminishing) rate by 900 min. The stronger, more pronounced cold pool was thus able to remain a formidable obstacle, helping to raise air parcels to saturation even though the environment from which they came was stabilizing. Also, while the rate at which the inflow environment was stabilizing was faster in the higher shear case, the storm was also propagating more quickly, and the progressive loss of convective instability during the mature phase may have been partially offset by the higher shear storm’s ability to draw in water vapor for condensation at a substantially more rapid rate.

It is clear that the effects of the Coriolis force on a model storm are complex, as are the interactions among the various constituent parts of a model storm itself. These topics will be addressed in the next two sections.

4. Discussion

a. Influence of the Coriolis force on the storm’s temporal behavior and spatial structure

A principal characteristic of all the nonrotational model storms we have made thus far is that those storms that survived long enough to reach the mature phase were able to attain stable configurations during that period. This was certainly true with respect to the widths of their active convection (precipitating) regions and approximately true for their trailing subsidence regions. The typical nonrotational model storm spread out quickly, modifying its surrounding environment over a very wide area, most dramatically and importantly on its rear side; but then sooner or later reached a point where it effectively ceased to expand, finding itself instead in a statistically steady state. During this
Fig. 13. Time-averaged fields for the moderate-to-high shear unbalanced Coriolis model storm for a period late in the simulation (around 900 min). Shown are (a) storm-relative across-line (east–west) horizontal wind (storm propagation speed is 19.8 m s\(^{-1}\) at this time), (b) along-line (north–south) horizontal wind and (c) perturbation water vapor mixing ratio. Contour intervals are 3 m s\(^{-1}\) in (a) and (b), and 0.7 g kg\(^{-1}\) in (c). The entire (untransformed) horizontal width of the model troposphere is presented; the stretched grid resides at \(|x| > 150 \text{ km}\) and the outline of the cold pool is superimposed on each panel.

apparently unending state, the typical storm evinced very little change in the intensities of its important characteristics or the scales of its most critical components. The storm was still periodically generating new convective updrafts and downdraughts that were swept rearward, and this resulted in a very slow but measurable increase in the storm’s cloud shield and a concomitant slight increase in the rearward extent of its trailing region during this time. However, the continued cell production and cloud shield expansion caused neither a significant accumulation of subsidence warming nor drying in the storm’s trailing regions, nor any important further modifications to the convective instability of its forward environment. Alterations to either would have prevented statistical steadiness from establishing or persisting.

In contrast, the typical Coriolis model storm was slower to expand outward when compared to its nonrotational counterpart, but continued to widen its zone of influence long after the control storm had reached its steady state. In addition, the intensity of the subsidence warming in the leading and trailing subsidence zones continued to increase. Both characteristics indicate that a steady state was unobtainable—at least not over typical model integration periods—when the Coriolis force was activated. Steadiness was undermined by the two major and related effects of rotation: the trapping of the subsidence warming due to storm-
produced gravity waves and the *braking* of storm-induced airflow perturbations. These two effects will be considered in turn.

1) **Subsidence Trapping**

The set of equations of the nonrotational model does not contain physically realistic mechanisms for effectively trapping or adjusting storm-produced internal gravity waves or modifying their horizontal propagation away from their convective region source. Physical diffusion associated with turbulence will indeed try to damp the waves but, far from the storm, it is *numerical forcings* associated with numerical diffusion (whether explicitly specified for stability or implicit in grid mesh designs or advection schemes), artificial boundary conditions, and other factors that determine the fate of propagating gravity waves. Indeed, the boundary conditions and the implicit components of diffusion are of paramount importance since it can be shown that typically chosen values of explicit “background” diffusion result in miniscule damping of gravity waves that thus require enormous horizontal distances to have serious effect (Bretherton 1987); particularly if the diffusion schemes used are of high order. On the other hand, artificial boundaries are well known to cause trapping and/or reflection of gravity waves and FO88 reported that this could have very serious impacts on the model results. Recall that FO89 found the small tendencies towards decay in their (nonrotational) model storms after very long integration times were reduced by the adoption of still wider, more gently stretched flanking grid zones.

The implementation of the Coriolis force, however, has introduced into the physical model a fundamental length scale for the model storm—the Rossby radius of deformation. This radius, first deduced from the geostrophic adjustment problem (cf. Gill 1982), is a function of the environmental stratification, the depth of the disturbance that generates the gravity waves, and the Coriolis parameter. Since the radius increases as the Coriolis parameter vanishes, it is infinite for the nonrotational case. When the Coriolis force is activated, however, the Rossby radius becomes a “straitjacket” of sorts, imposed upon the storm from without by virtue of the fact that the earth is rotating. For example, using a simple mathematical model, Bretherton (1987) demonstrated how compensating subsidence associated with a disturbance spread outward with time, at a gravity wave signal speed, until ultimately the warming filled a zone of a width determined by the Rossby radius when the disturbance was persistent enough.

In addition to being contained within a Rossby radius, Bretherton found that the subsiding motion was especially concentrated around the convective disturbance itself and exponentially decayed away from that point. He recognized that the confinement of the subsidence zone’s width concentrated the downwelling, thus producing more intense subsidence warming. This did occur in our rotational model simulations. The figures already presented show that the environmental response to the squall line convection expanded out slowly in time, to be ultimately limited roughly to within a Rossby radius (which Bretherton demonstrated is actually an ε-folding distance that is on the order of 300–1000 km or so in our application) of the central convective zone. Precise designation of a single Rossby radius is not possible since waves with a variety of wavelengths are involved and the excitor of the waves is far from being a well-defined point source.

Since the gravity waves were not permitted to propagate away endlessly, unmolested, the subsidence warming was strongly concentrated in the storm’s locality and accumulated with time. This is demonstrated in another fashion by Fig. 14a, which presents estimates of CAPE across a portion of the domain for the three mature phase-averaging intervals from the unbalanced Coriolis simulation. These estimates were computed from time-averaged temperature and moisture data using undiluted parcels assumed to be representative of conditions of the lowest 70 mb. For comparison, the CAPE of the initial, undisturbed environmental state is about 2500 J kg⁻¹.

Extremely low values of CAPE characterize the storm itself, where surface cooling and drying has combined with warming aloft to eliminate the convective instability with respect to a parcel of local, near-surface origin. Subsidence warming of the midtroposphere ahead of the storm has caused the CAPE decline from the initial state in that region; note the exponential decline of the response away from the leading edge as predicted by Bretherton (1987). Note also the general increase of the response with time as the subsidence warming has been forced to accumulate. The width of the low CAPE zone, a rough measure of the storm’s width comprising both the convective and trailing regions, has also increased with time, approximately doubling in size between 300 and 900 min, owing to the accumulation of trapped subsidence warming.

The remaining panels of the figure are presented in order to directly compare Coriolis simulations with their nonrotational control runs. Figures 14b,c compare CAPE estimates for Coriolis and control model storms—the lower (Δμ = 7.5) and higher (Δμ = 20) shear cases, respectively—made late in the simulations (around 900 min). For both cases, the effect of activating the Coriolis force was qualitatively similar. Compared to the Coriolis model storms, the widths of the low CAPE zones of the control storms were considerably narrower, owing to the lack of a mechanism for trapping and accumulating the subsidence warming. As a result, the subsidence responses in both the control storm’s forward and trailing environments were not only farther reaching but also much less concentrated than for the Coriolis storm since the rotationally-induced trapping effect was not in operation. The lack
Fig. 14. Figures showing horizontal distributions of convective available potential energy (CAPE in J kg⁻¹), with distance from domain center (km), for the central 2000 km of
the domain. Panel (a) presents time-averaged statistics from the low-moderate shear unbalanced Coriolis for the three averaging intervals previously analyzed, centered at 100 (solid
line), 300 (dashed line), and 900 (dash-dot line) km. Panel (b) shows the CAPE estimates from the mature phase for the unbalanced Coriolis and the model storms. Panel (c) compares the CAPE estimates from the mature phase for the low-moderate shear Coriolis (dash-dot line) and the model storms (solid line). The panel shows that the most significant changes occurred at a distance of 100 km, slightly faster than the domain translation speed.
of a subsidence trapping mechanism in the nonrotational storm has meant that there was a limit as to how much damage the storm could do to the convective instability of its own inflow.

Some subsidence trapping occurred in the control model storms, but the molestation of the gravity waves in those cases was virtually all numerical in nature. Figure 14d presents CAPE estimates from the lower shear nonrotational model storm at two instances in the mature phase, at 300 and 900 min. Recall that the control storm possessed a mature state characterized by statistical steadiness. Consistent with this steadiness, the width of the control storm was but little changed with time; some small expansion did occur, but the growth was quite small when compared to the Coriolis storm. Accumulation of the subsidence warming caused a decline in the CAPE in the far field of the control storm, but it is significant that the response was largest in those regions that were farthest from the central portion of the storm; while the storm's adjacent environment was virtually unaffected by the passage of a very long period of time.

Sensitivity tests revealed that the greater magnitude of the CAPE declines in the far field relative to the regions closest to the storm in the control simulation was a function of the stretching rate used in the flanking expanding meshes. Hence, the greater degradation of the far-field environment was spurious since it was numerically forced. As a gravity wave progressed through the stretched grid, it became increasingly difficult for the grid to resolve, and thus more of its energy was aliased into the larger scales. In addition, some amount of reflection was occurring in this zone due to the mesh expansion. The adoption of numerically wider, more gently stretched expanding meshes helped to alleviate these problems, actually by delaying them until the waves got farther from the central portion of the storm. FO89 determined that the effects of this artificial wave trapping in the far field was ultimately communicated back into the storm and was responsible for the very gradual decay tendency displayed at large integration times in the control model storm. However, it must be noted that the negative impacts of the grid design were felt only very late in the simulation and that the expanding mesh has behaved better than any other lateral boundary treatment we have tried.

One result of subsidence trapping, whether physically or numerically determined, was the spoilage of the originally favorable environment out ahead of the storm. As time passed, the environment that a Coriolis storm was advancing toward was having its convective instability eroded by midtropospheric subsidence warming, which was forced to accumulate locally to the storm due to the trapping (Fig. 14). Therefore, the storm was gradually poisoning itself with the waste products of its own convective engine. Although indirectly of the storm's own making, this effect, an obviously important contributor to the decaying trend, was imposed upon the storm from without by the alteration of the conditions of its forward environment.

2) Airflow braking

Another way that the rotational force appeared to be modulating (if not injuring) the squall line was through its braking of ground-relative wind perturbations. This is not truly separate from the topic of subsidence trapping since the braking effect of rotation is the mechanism that adjusts the gravity waves and produces the trapping. The Coriolis force provides a sink to across-line momentum produced by the storm, turning it clockwise into the orthogonal direction. Once an along-line wind component has been produced, it acts in opposition to the original across-line wind, again due to the Coriolis force. Therefore, in a sense, the storm's own convectively induced perturbations are used against itself.

The airflow perturbations that are acted upon by the rotational force are those in the ground-relative frame of reference; this is needed so that the entire storm system remains unaffected by reference frame translation (Galilean invariance). The importance of this lies in the fact that at all times during the mature phase, the storm's propagation speed (12.2 to 14.8 m s⁻¹ for the lower shear storm) was considerably greater than the fastest ground-relative wind at any level (7.5 m s⁻¹ above 2.5 km). This served to enhance the Coriolis-induced braking effect on storm-relative rear-to-front flows such as the rear inflow current and certainly contributed to the observed decline in intensity and rearward reach of the rear inflow current reported above. This was also true to a smaller degree for the higher shear case.

We noted earlier that an important role of the rear inflow current is to supply a continuous stream of dry air to the storm's cold pool in order to replace the pool air that is constantly being moistened by the evaporation of rainwater. While the rear inflow current provides the most efficient means for dry air to enter the pool, such air can also enter the storm from the forward side in between the cells. As the rear inflow declined with time in the Coriolis simulations, dry advection in from the front became relatively more important to maintaining the integrity of the cold pool. Indeed, by the end of the unbalanced lower shear simulation, the major source of the pool's dry air was from the front since the storm-relative rear inflow had degraded to the point where it could no longer import air across the cold pool boundary from the trailing region.

The amount of air that can enter the storm from the front side is a function of both the unsteadiness of the storm circulation (which permits the air to enter) and the midlevel storm-relative inflow (which pushes it in). If the storm circulation is not sufficiently unsteady, as in the very high shear cases of FO90, the updraft will present an insurmountable barrier to midlevel envi-
vironmental air, irrespective of the intensity of the storm-relative midlevel airflow. In the Coriolis simulations, the model storms were found to slow with time, which decreased the storm-relative inflow since the storms were moving faster than the winds at midlevels. At the same time, however, the storm tended to become more unsteady. Thus, the period between the production of new cells (and concomitant dry injections from the forward side) was systematically decreasing.

This tendency was most dramatically illustrated during the sudden transition that occurred in the $\Delta u = 7.5$ unbalanced Coriolis storm at about 550 min (see section 3b), which may be considered as a substantial increase in the storm’s unsteadiness. The post-transition storm consisted of a series of weaker cells produced with somewhat greater frequency. The maximum intensities attained by the storm’s updrafts were smaller after this time in part since each individual cell was cut off by its succeeding cell before it could intensify to the same degree as cells were able to before the transition, due to the increase in frequency. This transition apparently coincided with an abrupt shift in importance from the relatively steady rear inflow to the impulsive forward inflow in accomplishing the dry advection into the cold pool (and also the downward mass flux that compensates for the upward flux in the updraft as required by mass continuity). We have found that the transition, and attendant storm slowdown, was preceded by about 70 min by an equally sudden drop in the intensity of the rear inflow current, which then continued to degrade with time. The inefficiency of the forward-side dry-advection mechanism manifests itself in requiring interruptions in the general storm updraft, which results in a weakening of the storm immediately after the transition since individual cells were cut off from their low-level moisture source by the succeeding cell even earlier in their lifetimes. This prevents them from becoming as strong as they potentially could have been if the cell period had been longer.

b. Comparison with a simpler dynamical framework

The analytical investigation of Bretherton (1987) concerned convection forced by a maintained, surface-based heat source in a calm atmosphere set between two parallel plates. In a subsequent study, Bretherton (1988) set the heat source into motion in a fluid, unbounded from above. He found that both motion and the nonrigid upper boundary were important to the results. For example, the atmospheric response in the vicinity of the source grew logarithmically with time in the uncapped domain but reflection of gravity wave energy off the rigid lid forced a steady state. In fact, steadiness near the heat source was only obtained in the uncapped model when the Coriolis force was activated. These simulations suggest that the upper boundary condition may have a more profound influence on the current model results than previously surmised.

To test the sensitivity of the model storms to the rigid lid, several simulations were run with the Klemp–Durran (1983; hereafter KD) radiation upper boundary in a version of the compressible Klemp–Wilhelmson (1978; hereafter KW) cloud model. Using broadly similar numerical choices and physical conditions, the modified KW model has produced model storms comparable to the anelastic model to an acceptable degree. Storms grown in the rigidly capped and KD-topped domains differed solely in the stratosphere. Structural and behavioral discrepancies did not appear to be any larger when the Coriolis force was active. Each Coriolis storm still declined in intensity as subsidence warming collected in its vicinity and as its rear inflow current slackened. The KD boundary condition was built from a set of conditions that specifically neglected the Coriolis force, but we presently believe that a more sophisticated version would not alter these conclusions.

Just as there are many idealizations and simplifications in the model framework that restrain extrapolation of the results to real convective situations, there are numerous differences between the frameworks of the current study and the Bretherton experiments that inhibit direct comparison. For example, nonlinear effects were neglected in Bretherton (1987, 1988). His Boussinesq atmosphere possessed neither stratosphere nor viscosity and the heat source was not only artificially maintained but also ground-based. The lack of stratosphere and viscosity probably made the choice of the upper boundary more important.

We have made a series of simulations, using both models and upper boundaries, in which a surface heat source roughly comparable to Bretherton’s was placed into our more complex (but now shearless) environment. We have essentially replicated the atmospheric response deduced by Bretherton (1988), finding in the nonrotational case logarithmic growth in the intensity of the induced winds in the immediate vicinity of the heat source. This growth gave way to steadiness when the Coriolis force was included, again in agreement with the analytic results. Note that the discussion about steadiness or growth has focused on the heat source region, not the far field.

These simulations have given added confidence in the results of the squall line runs and point out that further research is necessary to explain the differences in atmospheric response between a squall line and the Bretherton heat source. Still, we can note that the heat source that best approximates a squall line would be elevated and probably tilted in the vertical as well. Preliminary work suggests that the atmospheric response is indeed quite sensitive to the elevation of the source; at least one nonrotational case displayed steadiness in the vicinity of the source rather than logarithmic growth after the source was raised. In addition, the complex, layered airflow associated with the squall line circulation is far different than that produced by the simple, ground-level heat source. Most important, however, is
the fact that the Bretherton heat source is externally maintained while the squall line is internally sustained. In the former, the adjustment is one-sided in that the important feedback between the atmosphere and the heat source is missing.

5. Synthesis

a. Remarks on Schubert et al. (1989)

With the present temperature, moisture, and low-level shear profiles, we have found that the implementation of a Coriolis force typical of midlatitudes has had a net deleterious effect on the strength, persistence, and longevity of the model storms. In each case, the Coriolis model storm was weaker when compared to its nonrotational counterpart. Also, while the nonrotational model storms were able to attain or nearly approach statistical steadiness during their extended mature phases, this wasn’t true of the Coriolis simulations examined herein. Instead of exhibiting steadiness, the strength of the Coriolis storms declined with time, albeit at relatively slow rates. Certainly, nothing in these simulations suggests that the Coriolis force is beneficial to a storm.

This conclusion stands in apparent contrast to a comment of Schubert et al. (1989), who examined the response of the atmosphere to a heat source (squall line) in a two-dimensional, semigeostrophic model on an $f$-plane. Their model supported a perturbation that spread out with time, filling a Rossby radius by 16 h and evincing a Coriolis-induced north–south wind field qualitatively similar to those produced in our models. Schubert et al. (1989) noted that one element of the atmospheric response was the destabilization of the upper troposphere behind the leading edge of the squall line, and remarked that this may play a role in the maintenance of the trailing stratiform rain region. They did not conclude that Coriolis effects were a favorable influence on the longevity of their model perturbation as a whole. No such conclusion could be made anyway since the heat source was externally maintained so that important feedback effects were again missing.

In the present model storms, the activation of the Coriolis force does result in a destabilization of the upper troposphere in the storm’s trailing region, produced by the combination of warming at middle levels from subsidence and cooling farther above, during the mature phase. Unlike the Schubert et al. experiment, the destabilization increases with time as the midlevel subsidence accumulates for reasons already discussed. Destabilization in general also occurs in nonrotational model storms, although for a given storm it is intensified in the rotational case due to its more substantial midlevel warming. While this may benefit the strength and longevity of stratiform precipitation (which requires ice microphysics to be simulated; see FO88), the Coriolis-induced degradation of convective instability in the forward environment unfavorably impacts the storm system as a whole and thus ultimately the stratiform portion of the storm as well.

b. Role of rear inflow in the storm’s decaying phase: RKW theory

In section 4, the influence of the Coriolis force was discussed in terms of two effects: subsidence trapping and airflow braking. The trapping effect led to the systematic decline in the convective instability of the inflow environment in the Coriolis model storms as well as the enhanced and continuing concentration of subsidence warming and drying in their trailing regions. The former tendency was clearly contributing to the slow weakening exhibited by these storms.

In contrast, the influence of the braking effect, which was most clearly manifested in the decline in the intensity and extent of the storm’s rear inflow currents, is less clear. On the one hand, the weakening of the rear inflow current may be participating with the subsidence trapping effect in injuring the model storm. We often observe that the strongest storms have the most intense rear inflows. This can be seen, for example, in the strictly 2D model storms of FO89 (see their Fig. 22), although it should be remembered that their rear inflows are exaggerated somewhat due to the model geometry’s confining the airflow to the across-line plane.

FO89 remarked that a stronger rear inflow could result in a more organized and deeper subcloud cold air pool. Intensifying the inflow current means driving more dry air from the trailing region into the pool, thereby enhancing the potential for evaporational cooling within the pool itself. The stronger airflow also tends, by increasing convergence, to concentrate the chilled pool air more narrowly in the region behind the pool nose, thereby promoting a stronger, deeper, more impermeable and formidable obstacle to the oncoming environmental air. The obstacle effect dynamically forces air parcels to rise, ideally until saturation is achieved and buoyancy forces can take over. This process may be particularly crucial in maintaining a storm under marginal or stabilizing conditions if the cold pool obstacle can remain sufficiently deep and impermeable, as has appeared to be the case in the moderate-to-high shear Coriolis model storm. Further, FO89 noted that the intense rear inflows in their higher shear cases seemed to be “propping up” their storm updrafts since they were so elevated and their cold pools were so deep.

On the other hand, the intensification of the storm’s cold pool that can result from a strengthening rear inflow may injure a storm. If this is true, then it seems logical that if a rear inflow current and cold pool were to decline in strength with time, as found in the Coriolis model storms, it would help to at least partially counter any other storm weakening effect in operation. This argument stems from an application of Rotunno et
al.’s (1988; hereafter RKW) theory to our simulations. RKW argued that for a given amount of low-level wind shear there is an optimal cold pool strength where the inherently negative influences of both the cold pool and the shear balance to produce a net positive effect on the storm. This balance, the “optimal state,” results in the environmental airflow that impinges on the cold pool nose being turned upward into the strongest, most vertically oriented storm updraft possible. The situation in which the shear is stronger than the cold pool, the “superoptimal state,” is one in which the updraft is impelled to tilt in the downshear direction with height, forcing it to rain into its own inflow. When the cold pool dominates, the “suboptimal state,” the storm updraft leans upshear. FO89 identified the suboptimal state to be the typical condition of multicell storms, which are characterized by a sequence of individual cell updrafts embedded in a general upshear tilting airflow.

In each Coriolis model storm, the depth and intensity of its cold pool clearly declined with time. Concurrent with this, the low-level shear of the inflow environment, even in the immediate vicinity of the storm, did not change very much. Therefore, the storm, which was clearly in a suboptimal condition throughout its mature phase (the general storm updraft above the cold pool was considerably tilted in the rear or upshear direction), should have been approaching the optimal state with time. Thus, all other things being equal, the storm should have been strengthening during the course of its mature phase.

All other things were clearly not equal however, as the storm still had to contend with the decreasing convective instability of its inflow environment. However, the storm’s weakening should not have sprung from its updraft airflow becoming more tilted (more suboptimal) with time. Indeed, Fig. 5 suggests that the weaker shear Coriolis storm was resisting becoming more tilted during its decaying stage. The intensity of the storm-relative airflow clearly weakened between 300 and 900 min [panels (a) and (c)], but there is no indication that the general orientation of the storm’s time-averaged updraft became substantially more tilted from the vertical during that interval; an observation corroborated by the storm-relative vector airflow fields (not shown).

In this and the other Coriolis model storms that have been examined however, any potentially positive benefit the suboptimal storm may have gained from the weakening of its cold pool had to be overshadowed by the negative effects of subsidence accumulation in the storm’s locality. The airflow braking effect, even if it were in opposition to the decaying trend, was insufficiently strong to prevent the model storms from exhibiting decay. If RKW’s theory is substantially correct and applicable to the current situation, as it is (see below), one could see how the decline of the rear inflow current would help to moderate (if not reverse) the decaying tendency of a Coriolis model storm. For example, if the rear inflow of a storm were to remain strong and firmly rooted in the trailing region, then as the storm aged it would be accessing air that was progressively drier due to the trapping effect, which would in turn make the cold pool even stronger and increase the storm’s propagation speed. In that event, the storm would be observed to accelerate as it weakened at a (perhaps substantially) greater clip, with negative influences owing to subsidence trapping and increasing suboptimality in concert.

c. Further comments on RKW theory

In concluding this discussion, we are compelled to express caution about the application of RKW theory. The analysis in RKW was based on a simplified physical framework which, as a first approximation, did not attempt to address the possibility of storm-relative airflow within the cold pool itself. However, observational and modeling work have demonstrated that there are complex and potentially important air motions in this zone; namely, the rear inflow current and the front-to-rear return flow(s) (see FO89’s Fig. 27). It has been recognized that these air currents are indispensable parts of the quasi-equilibrium state in the mature phases of nonrotational storms.

For this reason, we have been concerned with the neglect of this complex cold-pool airflow in the derivation of RKW’s theory, particularly since it has become common to cite and employ their simple balance equation for the optimal state. This equation is written as

$$\Delta u = c,$$  \hspace{1cm} (1)

where $\Delta u$, the wind speed change over the cold pool depth, and $c$, the propagation speed of an idealized density current, are proxies for the positive vorticity associated with the environmental wind shear and the negative vorticity produced within the cold pool, respectively. Equation (1) was derived by integrating the horizontal vorticity equation around a specified control volume enclosing the vertical interface between the cold pool and the inflow environment.

Among the assumptions used in the derivation was the specification that there was to be no inflow into the storm from the front at midlevels and that there was no (storm-relative) motion within the cold pool itself. The former appears to be an acceptable approximation, if somewhat restrictive; for example, FO89 found that for their very strongest model storms, which were those with the most intense low-level wind shear intensities, there was very little midlevel inflow since their storm speeds closely matched the ground-relative winds above the shear layer. A similar tendency can be noted in the 3D simulations of Weisman et al. (1988). Applicable midlevel inflow, as found in the lower shear cases, may injure a model storm’s general updraft by exerting a
rearward push on it and by forcing its dilution with dry, midlevel air.

The problem at hand is not that an intensifying, rear inflow current produces, through enhanced dry advection, a colder subcloud air pool than could exist if the current were weaker, as discussed in subsection 5b. That situation is implicitly included in the RKW framework since the theory is specifically concerned with the cold pool intensity as it exists, not with how it was created or is maintained. Rather, it is the possibility that the subcloud airflow may substantially alter the magnitude of the cold pool’s negative vorticity through vorticity advection that is of concern. We noted above that a rotational storm’s weakening rear inflow current may have been trying to oppose the general decay tendency imposed upon the storm by other forces, since it was tending to reduce the negative vorticity of an overly strong cold pool that was making the storm system suboptimal. If at the same time the net vorticity advection by the pool’s airflow were either negligible or acting in concert with its implicit effects, then there is no problem with the above discussion. However, if the vorticity advection is in opposition, such that it mitigates or even reverses the implicit effects, then the current application of RKW theory is suspect.

Unfortunately, we are not presently in a position to draw firm conclusions about this topic. However, we should like to provide to the debate a few points gleaned from current modeling work. For example, one might suppose that the effect of rear inflow current feature on the vorticity balance of the cold pool increases as the current itself intensifies, so that RKW’s theory as it stands is least applicable when the rear inflow becomes strong. There is evidence to dispute this supposition since, at least in 2D, the rear inflow current cannot be considered separately from its associated return flows that must be present to satisfy mass continuity. In the F089 study, the highest shear storms appeared to be closest to optimal in that they possessed the most intense and vertically oriented storm updrafts. Values of $c$ for these cases satisfied (1) to a high degree; in the largest shear case, for example, $\Delta u = 30$ and $c$ was estimated to be between 30.1 and 31.4. Yet, those same high shear cases also had the most intense storm-relative rear inflow currents. This indicates that either the airflow within the pool doesn’t matter at the optimal rate or that the net effect of its components (rear inflow and return flows) is zero due to cancellation.

The latter is the more likely scenario, a belief borne out by calculations. Therefore, whatever the influence of the cold-pool airflow, for good or ill, it apparently diminishes as the storm approaches the optimal state, at least in the 2D model simulations. It remains then to discern whether the influence of the subcloud airflow on the suboptimal storm is beneficial or detrimental; that is, given a suboptimal state to begin with, would enhancing or rearranging the rear inflow and return flow(s) benefit or injure the storm? An analysis of this question would have to carefully reconsider more of the assumptions and idealizations RKW was justified in making for the optimal case since they would no longer apply. Two such simplifications were that the storm-relative inflow above the cold pool from the front side is zero and that the integral of the vertical flux of vorticity through the top of the analysis domain vanishes. The former is invalid for a suboptimal storm since they can experience considerable relative midlevel inflow and the latter is only clearly true for a vertically oriented, symmetric updraft which, since it lacks tilt, is by definition not suboptimal. In addition, the even more fundamental approximation of the horizontal vorticity $\eta (=-\partial u/\partial z - \partial w/\partial x)$ as $\partial u/\partial z$ along the left lateral boundary of the control volume is problematic if the air in the storm’s tilted front-to-rear airflow is still rising when it crosses this boundary.

We eagerly await work to be done in this direction. When completed, an important piece of the puzzle will have been found and we will know what influence on the life cycle of storms an evolving, time-dependent mature phase subcloud airflow may have. Such time-dependence may be induced by forces other than the Coriolis force, which might act more strongly or quickly on the storm than rotational effects were able to, but have not yet been incorporated into the increasingly sophisticated but still quite simple model framework.

6. Summary and conclusions

Described herein is the temporal behavior and spatial structure of two-dimensional storms grown in a model framework that included Coriolis forcing appropriate for midlatitudes. This work extends two previous studies (Fovell and Ogura 1988, 1989) that examined model storms grown under similar environmental conditions in nonrotational, strictly two-dimensional settings. The simulations examined in greatest detail were initialized with a low-level wind shear profile of low-to-moderate intensity. Two Coriolis simulations with this shear were described; both included the effect of Coriolis accelerations on the momentum equations but only one specifically incorporated the impact of an assumed along-line temperature gradient geostrophically balanced with the vertical wind shear. Sensitivity to the specific intensity of the shear was discussed through comparison of these results with those from a higher shear case.

Unlike their nonrotational counterparts, none of the Coriolis model storms made were able to attain or maintain what we have previously referred to as a “quasi-equilibrium” state. Quasi-equilibrium storms possess mature phases characterized by essentially statistically steady behavior with respect to storm strength, propagation speed, and other influential characteristics when considered over periods longer than those of the individual cells. Instead, the Coriolis storms possessed
mature phases marked by gradual but definite decay in the production of condensation and precipitation and the slowing in storm speed. These are the first model storms created with the present model sounding and wind profiles that have terminal mature phases brought about as a result of physically realistic forcings. However, the time scale of the decay, at least in these cases, makes it unlikely that Coriolis forcing is the primary mechanism behind the demise of real squall line thunderstorms, although it may play a secondary role.

In each rotational case, the decay phase was marked by two major temporal trends that were absent in the nonrotational simulations: the gradual contamination of the forward environment with storm-induced subsidence warming and the decline in intensity of the rear inflow current. These two trends were analyzed in terms of the two major influences of the Coriolis force on model storms, which were termed "subsidence trapping" and "airflow braking." The addition of Coriolis accelerations to the model framework provided a physical mechanism for limiting the spread of storm-induced downwelling away from the convective region, causing the subsiding motion to become trapped in the storm's vicinity. As a result, subsidence warming and drying in a Coriolis model storm was stronger and more concentrated in the storm's locality than was found to occur in its nonrotational counterpart, and also continued to accumulate during the decaying phase. This slowly eradicated the convective instability of the air flowing into the storm and certainly played a major role in the storm's slow decay. Subsidence also built up on the rear side of the storm, producing progressively warmer and drier air in the trailing region that made the storm appear to continue to expand farther rearward with time. In contrast, the nonrotational model storms were statistically steady with respect to their most important features, including the temperature and moisture conditions in the storm's enveloping environment, during maturity.

Airflow braking was the term used to describe the Coriolis force's turning action on storm-induced airflow perturbations, which removed momentum from one horizontal direction and deposited it into the other. The braking effect acted most strongly on the largest (ground-relative) airflow perturbations, the most important of which was the rear inflow current. This was why the rear inflow current degraded in intensity and horizontal extent with time as the storm aged. This current is an important source of dry air for the cold pool, which uses it to replace air that has been moistened by the evaporation of rainwater entering the pool from the cloud above. Dry advection into the cold pool was diminishing with time in all of the Coriolis simulations due to the progressive weakening of the inflow current. In the lower shear cases, the rear inflow was even losing its root in the trailing region, which meant that dry air importation into the cold pool from the rear effectively ceased near the end of the simulation.

Dry air can also be obtained from the front side of the storm, due to the unsteadiness of the circulation; but this process appears to be less efficient. The net decline of dry advection caused the gradual decline in the size, intensity, and dryness of the cold pool, which depends on a maintained flow of dry air to sustain itself.

The accumulation of subsidence warming clearly injured the model storm. The role that the declining rear inflow played in the decay phase is less clear and requires additional study. On the one hand, a strong inflow current helps to deepen the cold pool so that it presents a more formidable obstacle to the oncoming airflow. On the other hand, the recently proposed theory of Rotunno et al. (1988) suggests that the resulting stronger pool would force the already suboptimal storm to be farther from an optimal condition, rendering the storm less intense—so that the weakening rear inflow should by itself have a positive effect. Thus, the decaying tendency may represent a net, negative balance result of two opposing effects.

It was found that the inclusion of the geostrophically balanced along-line temperature gradient had small but measurable consequences in this situation. Warm advection at low levels ahead of the storm tried to negate the effect of warm advection aloft on the convective instability, and cold advection into the cold pool opposed the general decline in pool intensity. The net effect was that the rotationally forced, mature-phase decaying tendency was slowed somewhat, but not arrested. However, it is possible to find situations in which the geostrophic balance term would be of greater relative importance, in both the forward environment and the subcloud pool. Also, it is also possible to design shear profiles that would cause the geostrophic advection in the forward environment to actually hasten the decline of convective instability there; for example, putting a shallow, shearless layer near the surface would induce a low-level temperature inversion to develop.

The more highly sheared storm evinced qualitatively similar responses to the rotational force. In the weaker shear case however, the influence of the Coriolis force was mainly felt behind the storm's leading edge; although we have seen that its impact on the forward environment was of crucial importance. In comparison, the stronger shear case showed that the atmospheric adjustment need not be so biased to the rearward side.

We have written solely of net negative influences for the Coriolis force. Indeed, in no case produced thus far have we seen the Coriolis force to play a positive role in the strength or longevity of the model storms. It has been commented (Schubert et al. 1989) that due to its destabilizing effect on the upper troposphere of the trailing region, the Coriolis force may have a preservation effect on the stratiform precipitation region. It may even be expected that a storm could become rotationally stabilized, and thus rebound in strength after one inertial cycle. Yet, these simulations show,
within the limitations of two-dimensionality, that the rotational force has a deleterious influence on the storm system as a whole. With rotation activated, its trapping effect causes the subsidence warming, the waste exhaust of the model storms, to continually accumulate in the storm's locality, and thus the storm fouls its own nest with poisons of its own manufacture.

Acknowledgments. The author expresses his appreciation to the following individuals for aiding the author in his research and/or reviewing the manuscript: Drs. C. Bretherton, M. I. Biggerstaff, J. R. Holton, R. A. Houze, Jr., and E. J. Zipser. The author also appreciates the input of the journal reviewers and, especially, the editor for his interest and comments. This work was supported by the National Science Foundation under Grants ATM-8719838 and ATM-8914852. The numerical simulations were made using the Cray X-MP computer of the National Center for Supercomputing Applications at the University of Illinois at Urbana-Champaign.

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